

**Review Comments for “Snow-darkening versus direct radiative effects of mineral dust aerosol on the Indian summer monsoon: role of the Tibetan Plateau” by Shi et al.**

The authors conducted a set of GCM simulations to quantify dust SDE and DRE over the Tibetan Plateau and its impacts on Indian monsoon onset. They found that dust SDE and DRE exert opposite effects on Indian monsoon onset and proposed a possible mechanism. The results are interesting and the authors did a generally good job in writing the manuscript. However, some parts of the manuscript still need to be improved, particularly for model descriptions and evaluations. Please see my following comments.

RE: Thanks for the positive comments.

Major Comments:

1. Section 2 (Model and Experiments): Since this work is a modeling study, the model descriptions require more details. Here are some examples. (1) What is the new dust size distribution used in CAM4-BAM? How many size bins are used and what are the values for these bins?

RE: There are four size bins (0-1.0  $\mu\text{m}$ ; 1.0-2.5  $\mu\text{m}$ ; 2.5-5.0  $\mu\text{m}$ ; 5.0-10.0  $\mu\text{m}$ ) in diameters used in this study. In this version, the percentages of emission (0.02, 0.09, 0.27, 0.62) is modified to allow more large particles.

We noted this in the revision (Page 3 Lines 28-32): “The dust cycle including the emission, transport and deposition, is parameterized in CAM4 and its radiative feedbacks are also calculated on line. The dust sizes in CAM4 contain four bins of 0-1.0  $\mu\text{m}$ ; 1.0-2.5  $\mu\text{m}$ ; 2.5-5.0  $\mu\text{m}$ ; 5.0-10.0  $\mu\text{m}$  in diameters, respectively (Mahowald et al., 2006). The CAM4-BAM has been improved by an optimized soil erodibility map and a new size distribution for dust emission (the percentages for four bins are 0.02,0.09,0.27,0.62, respectively), as well as updated optical properties for radiation budget, to present a better performance on simulating the global dust cycle (Albani et al., 2014)”

(2) For dust optical properties, what have been updated?

RE: Several optics in the CAM4-BAM were improved, as introduced in Page 7 of Albani et al., (2014). Due to limited space, we did not introduce these changes in our paper. We referred to Albani et al.’ paper for detailed information (Page 3 Lines 31-32).

(3) The model simulations did not include aerosol indirect effect but used prescribed CCN. Does this mean that aerosol wet removal through in-cloud process is not included? If so, this could cause large uncertainty in simulations. Please clarify how

the prescribed aerosol in-cloud process would affect aerosol wet deposition.

RE: The reason we chose the model CAM4 not CAM5 is that CAM4 does not include aerosol indirect effect, which is complicated with large uncertainty and not our focus. Certainly, the choice of CAM4 leads to other uncertainty, e.g., the in cloud removal of dust, as the reviewer pointed out. We admit this bias in simulations, however, the bias in wet deposition does not affect our discussion because wet deposition occupies a small part of total deposition over dust source regions.

We emphasized this possible bias in the revision (Page 3 Lines 32-33; Page 4 Lines 1-2): “In CAM4-BAM, the SDE of all aerosols are enabled but the indirect effect is not considered, which means that the aerosol changes in cloud process as condensation nuclei are prescribed. Wet removal through in-cloud process is not considered, which may induce bias of dust deposition on snow over Asia.”

(4) The authors used the SNICAR model to deal with snow darkening processes. How does the model handle aerosol-snow interactions? To my understanding, SNICAR assumes external mixing between aerosols and spherical snow grains. But recent studies have suggested that aerosol-snow internal mixing and nonspherical snow shape could significantly affect aerosol-induced snow albedo effects (e.g., Flanner et al., 2012; Liou et al., 2014; Räisänen et al., 2017; He et al., 2018), which may introduce some uncertainty in the simulations here.

RE: Thanks for the comment. SCINAR assumes external mixing between aerosols and spherical snow grains, which may induce uncertainty.

We referred the publications and discussed the uncertainty of simplification of spherical snow grains in this model (Page 4 Lines 7-9): “Of note is that SCINAR assumes external mixing between aerosols and spherical snow grains, however, aerosol-snow internal mixing and nonspherical snow shape could significantly affect aerosol-induced snow albedo effects, based on recent studies (Flanner et al., 2012; Liou et al., 2014; Räisänen et al., 2017; He et al., 2018).”

(5) How does the model deal with the aerosol removal in snowpack? Does it assume a fixed removal efficiency?

RE: Yes. In the model, it assume a fixed removal efficiency, that is, the removal by meltwater is proportional to its mass mixing ration (multiplied by a scavenging factor).

(6) The way to calculate SDE and DRE by computing the difference between EXP1 and EXP2 and between EXP2 and EXP3 has an underlying assumption that SDE and DRE are linearly additive. However, SDE and DRE could have interactive and nonlinear effects, which makes the calculations above inaccurate. For example, if we

refer EXP4 to a new experiment with only SDE enabled, then how different would the result be if calculating DRE by taking the difference between EXP1 and EXP4, compared with “EXP2 minus EXP3”. And how different would the result be if calculating SDE by taking the difference between EXP4 and EXP3, compared with “EXP1 minus EXP2”. Do the authors have any suggestions on which way of calculation is more accurate in terms of quantifying dust SDE and DRE?

RE: Yes, the reviewer is right. The nonlinear term indeed exists because of the interaction between two effects. A good way to examine the nonlinear term is to conduct a fourth experiment with only SDE enabled and see whether EXP1-EXP4 and EXP2 and EXP3 is consistent (if yes, it means the nonlinear term can be neglected). In this study, we mainly focused the SDE of dust so we used EXP1-EXP2 (not EXP4-EXP3) because we considered the EXP1 as control experiment and in real world these two effects are indeed enabled. However, due to limited time for final response phase, it is difficult to finish the EXP4 to examine the nonlinear term (We have already added three more experiments on black carbon. Please see the response for RC2). If the reviewer insists, we wish to finish the EXP4 in next phase.

#### References:

- Flanner, M. G., et al.: Enhanced solar energy absorption by internally-mixed black carbon in snow grains, *Atmos. Chem. Phys.*, 12(10), 4699–4721, 2012.
- He, C., et al.: Impact of grain shape and multiple black carbon internal mixing on snow albedo: Parameterization and radiative effect analysis, *J. Geophys. Res.-Atmos.*, 123, 1253–1268, 2018.
- Liou, K. N., et al.: Stochastic parameterization for light absorption by internally mixed BC/dust in snow grains for application to climate models, *J. Geophys. Res.-Atmos.*, 119, 7616–7632, 2014.
- Räisänen, P., et al.: Effects of snow grain shape on climate simulations: Sensitivity tests with the Norwegian Earth System Model, *The Cryosphere*, 2017.

2. Section 3.1 (Model validation): (1) For the AOD evaluation, since the model simulation did not include non-dust aerosols, it is not an apple-to-apple comparison for modeled and MISR AOD here. The AOD comparison did not give us very useful information. If the authors want to use total AOD from observations, the model simulations need to include all aerosol types. If the authors only want dust AOD, maybe CALIPSO observations could help. Focusing on AOD over dust source regions can also be a way to evaluate modeled dust AOD, but in that case it is difficult to know how the model performs in terms of dust transport, particularly over remote regions such as the Tibetan Plateau. Besides, even over the dust source regions such as north of Tibetan Plateau, the modeled AOD is much smaller than MISR AOD. What would be the possible reasons?

RE: Thanks. We used the CALIPSO data instead in the revision. Compared to the dust

AOD in CALIPSO data, the simulated AOD is smaller, especially over the source areas (Figure R1). The possible reasons are as follows. Over the dust source, there are lots of dust with larger particle sizes but in the model, the considered dust particles are restricted to less than  $10\mu\text{m}$ . Thus, the dust forcing is underestimated due to less coarser dusts in the current global climate models (Kok et al., 2017). Also, dust model may have quite large differences in simulating vertical distribution, emission, deposition, and surface concentration of dust (Pu and Ginoux, 2018), which affects the AOD as an integrated variable. Furthermore, the spatial resolution of model is not fine enough, which fails to well resolve the complex topography and dust sources over East Asia.

We mentioned these in the revision (Page 5 Lines 5-8): “The simulated absolute values of dust AOD over Arabian Peninsular, southwestern slope of the TP and Taklimakan desert are biased low because the considered dust particles are restricted to less than  $10.0\mu\text{m}$  and the dust forcing is underestimated due to less coarser dusts in the current global climate models (Kok et al., 2017). ”

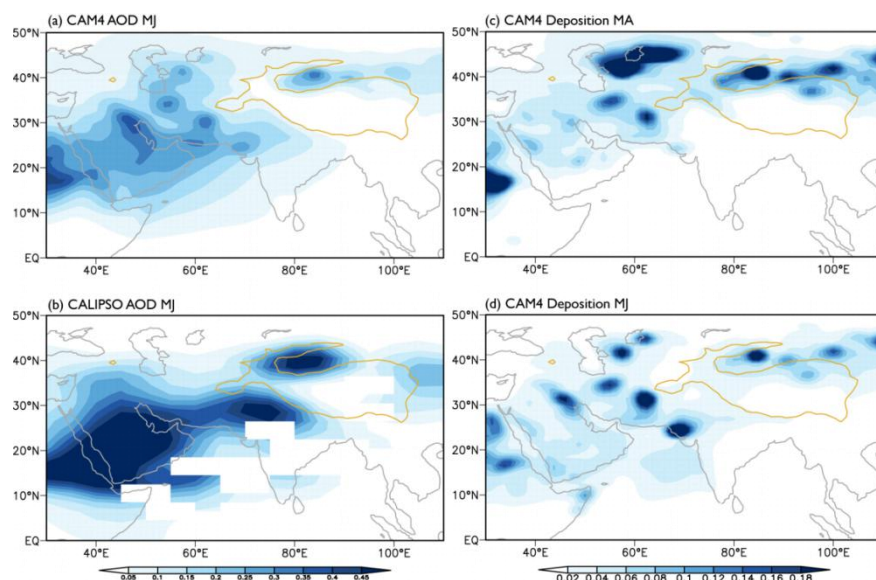


Figure R1: Averaged dust aerosol optical depth over Asia for May and June in CAM4 (a) and in Cloud-Aerosol Lidar and Infrared Pathfinder Satellite Observation (CALIPSO)-retrieved data for 2007-2011 (b); and dust deposition flux including both dry and wet deposition for March and April (c) and for May and June (d).

(2) Also it seems that MODIS AOD is better than MISR AOD at least over dust source regions, due to the MODIS deep blue retrieval algorithm. Why did the authors select MISR instead of MODIS?

RE: See the response above. We used CALIPSO data for dust because it can be directly compared to our simulated AOD.

(3) The authors described in detail the consistency and inconsistency between model

simulations and observations in terms of AOD, snow cover, and monsoon climatology, but it appears that not enough explanations have been provided for the model observation differences. The readers may also want to know the reasons causing the model observation discrepancies, which would be very useful for future model improvements.

RE: For the reasons for AOD differences, we discussed it in previous response 2-(1). The main cause for the underestimation on snow cover fraction and precipitation over monsoon regions is obviously the coarser resolution in our model. For example, our model can not resolve the high mountains over Tianshan mountains and western Tibetan Plateau, which the simulated snow cover fraction is relative smaller. Similarly, thin high topography over western edges of Indian subcontinent and Indo-China Peninsula is also partly missed so the observed maximal precipitation at these regions are also smaller in model.

In the revision, we added these explanations (Page 5 Lines 5-8, 17-20): “The simulated absolute values of dust AOD over Arabian Peninsular, southwestern slope of the TP and Taklimakan desert are biased low because the considered dust particles are restricted to less than 10.0  $\mu\text{m}$  and the dust forcing is underestimated due to less coarser dusts in the current global climate models (Kok et al., 2017).”, “Over the western TP, the MODIS observation presents a fraction larger than 80% but the simulated fraction is smaller. In particular, the model underestimates the elevations of finer-scale mountains and corresponding snow cover fractions due to the coarser resolution, e.g., over the Tianshan mountains.”

(4) Since the snow darkening effect (i.e., albedo reduction) is one focus in this work, it would be straightforward to consider evaluating modeled snow/surface albedo at least over the Tibetan Plateau, for example, by comparing with MODIS albedo product. Is there any specific reason for the authors to leave out this part?

RE: Thanks. We did not show the albedo because the albedo is directly controlled by the snow cover fraction. Compared to the MODIS data over the TP (Meng et al., 2018), the model captures its spatial distribution but overestimates the surface albedo, which is similar with multi-model ensembles (Li et al., 2016), mainly due to the overestimated snow cover fractions.

Minor Comments:

1. Page 1, Line 16: I suggest replacing “clarified” with “quantified”.

RE: We kept unchanged because we can not quantify their links.

2. Page 2, Line 10: Please remove “reflect,” since reflection is part of scattering.

RE: Removed.

3. Page 2, Lines 31-34: For the authors' information, some recent studies on BC/dust SDE are missing here, which improved the understanding of aerosol SDE particularly over the Tibetan Plateau. Some examples are listed as follows.

References:

He, C., et al: Black carbon radiative forcing over the Tibetan Plateau, *Geophys. Res. Lett.*, 41, 7806– 7813, 2014.

Zhao, C., et al.: Simulating black carbon and dust and their radiative forcing in seasonal snow: a case study over North China with field campaign measurements, *Atmos. Chem. Phys.*, 14, 11475-11491, 2014.

Lee, W.-L., et al.: Impact of absorbing aerosol deposition on snow albedo reduction over the southern Tibetan plateau based on satellite observations, *Theor. Appl. Climatol.*, 129(3-4), 1373-1382, 2017.

Niu, H.W., et al.: Distribution of light-absorbing impurities in snow of glacier on Mt. Yulong, southeastern Tibetan Plateau, *Atmos. Res.*, 197, 474-484, 2017.

RE: Thanks for the references and we added them (Page 3 Line 4).

4. Section 1 (Introduction): It seems that the authors did not mention their motivation to focus on the Tibetan region particularly. Thus, I suggest adding a short paragraph to highlight the importance of Tibetan Plateau (such as its role in altering Asian water resources and hydrological cycle), although the authors already mentioned a little bit in the descriptions of dust effects.

RE: Following the comments of another reviewer, we largely reorganized the results and discussion in the revision. In the original manuscript, the emphasized role of Tibetan Plateau may be not appropriate (the reasons can be detailed introduced in response RC2). Thus, we do not add a paragraph to introduce the role of TP.

5. Page 9, Line 6: please remove “is” before “occurs”.

RE: Removed.



**Review of “Snow-darkening versus direct radiative effects of mineral dust aerosol on the Indian summer monsoon: role of the Tibetan Plateau” by Shi et al.**

This paper examines the dust snow-darkening (SDE) and direct radiative effects (DRE) on Indian summer monsoon (ISM) with global climate model simulations. The authors found that dust SDE (DRE) tends to induce a warming (cooling) over Tibetan Plateau, and weakens (intensifies) the ISM. The main findings of this manuscript are contradictory to previous studies, but the authors did not provide convincing explanations. Thus, this manuscript needs careful revisions to meet the standard of Atmospheric Chemistry and Physics and resubmitted.

RE: Thanks for the comments. The reviewer said that the main findings of our paper are contradictory to previous studies. But we do not agree with the reviewer's viewpoint although we admit that some inaccurate arguments in the original manuscript may mislead the readers. We overemphasized the role of Tibetan Plateau in the original manuscript and it is actually not accurate based on new experiments and results.

We agree with the reviewer that we did not provide convincing explanations previously, thus, we conducted three additional experiments, with a special focus on black carbon. The aim is to examine whether the SDE and DRE of black carbon is similar with mineral dust or not. In most previous studies, black carbon is mainly considered to understand the effect of absorbing aerosols but mineral dust is indeed different from black carbon both for spatial distribution and radiative effect. Fortunately, the black carbon experiments tell that the SDE and DRE both intensify the Indian summer monsoon during the onset, which are in good agreements with previous studies (e.g., Lau et al., 2006; Qian et al., 2011).

The SDE of black carbon warms the surface over Tibetan Plateau (TP) and intensifies the monsoon during the onset, consistent with what were found in Qian et al (2011). Interestingly, the same model with the SDE of dust gives a quite different response of monsoon, which indicates different mechanisms behind SDE of dust and black carbon. As we proposed, the spatial distributions of dust and black carbon are not similar. The main difference is that black carbon from the industrial countries is generally transport eastwards and scarcely into upwind Central Asia. Central Asia is also covered by snow although far less than TP. As a result, the forcing of black carbon is restricted to TP but the forcing of dust is over Central Asia and TP. Westwards/northwestwards expansion of warming also shifts the pattern of low level circulation change, which weakens the summer monsoon.

The DRE of black carbon also warms the surface over the TP and intensifies the monsoon during the onset, consistent with Lau et al. (2006). Although the 3D distribution of black carbon are not the same, our experiments also support the DRE of black carbon can strengthen the monsoon. Comparing the responses to dust and black carbon under the same model and experiment design, we found that the summer monsoon during the onset is both intensified no matter whether it is warming or cooling over the TP. As the reviewer said, the TP cooling is unlikely to intensify the

monsoon (at least few evidence support it). We agree it and ascribe the monsoon strengthening to the warming over Arabian Peninsula/Middle East, which also gains support from previous researches (Vinoj et al., 2014; Jin et al., 2014; Solomon et al., 2015). Although the TP cooling tends to weaken the monsoon, the Middle East warming overacts and induces a stronger monsoon instead. In addition, the simulated TP cooling might be model dependent because DRE of dust is largely uncertain and depends closely on the size distributions, optical properties and etc (Kok et al., 2017). Anyway, our results support the important role of Middle East warming.

To summarize, we found different mechanisms for Indian monsoon to dust and black carbon forcing. The significant contributions from temperature changes over source areas (Central Asia for SDE and Middle East for DRE, respectively) are highlighted. The role of TP we proposed previously is not accurate. In the revision, we removed it and changed to “role of dust source temperature changes”. Detailed results and responses are shown in the following. We wish the current version of manuscript could give more convincing results and arguments.

Major comments:

The Indian summer monsoon is primarily driven by the thermal contrast between land and ocean (Wang et al., 2000). The up troposphere meridional temperature gradient south of Tibetan Plateau is one of the key controls of the Indian summer monsoon (Li and Yanai, 1996). An up troposphere warming over TP tends to increase the meridional temperature gradient, and intensifies the Indian summer monsoon (Wu et al., 2005; Liu et al., 2001). Previous studies show that both DRE and SDE of absorbing aerosols could induce a warming around TP and intensify the Indian summer monsoon (EHP effect), which is in general consistent with the observed relationship. In this study, however, the authors found the Indian summer monsoon is intensified (weakened) associated with cooling (warming) over TP, which is just opposite to previous studies. The authors should carefully check the model settings and give some explanations.

RE: Please see the next two comments for detailed explanations.

The authors found that the dust SDE induced TP warming tends to weakens Indian summer monsoon, which is opposite to what found in Qian et al. 2011. The authors stated that the opposite response is due to the difference in TP warming center distribution. Lau et al. 2010 found that aerosol SDE could produce an “elevated-heat-pump (EHP)” effect and increase the precipitation over Indian in May. Their results is in generally consistent with Qian et al. 2011, although the warming center is over western TP. The authors should provide convincing explanations why the response to dust SDE in this study is contradictory to previous studies.

RE: Yes. From previous studies (e.g., Lau et al., 2010; Qian et al., 2011), it can be obtained that the SDE of black carbon or all absorbing aerosols intensifies the Indian monsoon, which is different from what we simulated for dust. We pointed out that our



study is not contradictory to previous studies because our study merely focused on dust and the spatial distribution of dust is different. In our opinion, the key point is that we can not directly compare our results and previous studies. Thus, we conducted three additional experiments, with a special focus on black carbon, to support our study. The aim is to examine whether the SDE and DRE of black carbon is similar with mineral dust or not.

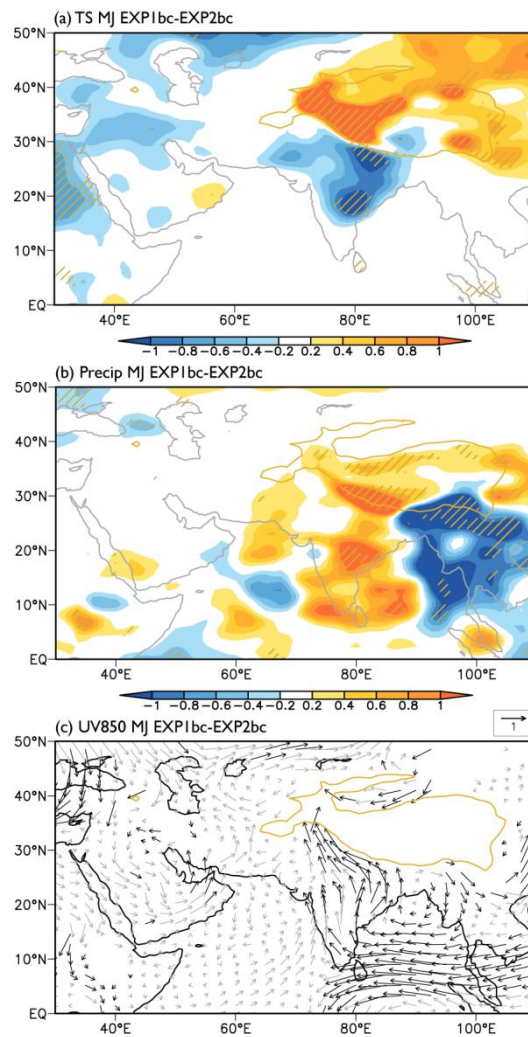


Figure R2: Spatial distribution of changes in precipitation rates (a,  $\text{mm day}^{-1}$ ), surface temperature (b) and 850hPa wind vectors (c, m/s) in May and June induced by snow-darkening effect of black carbon.

The SDE of black carbon warms the surface over western TP only, with a cooling over northern India (Figure R2a). The southerly winds and precipitation over India are significantly larger (Figure R2b, R2c), which indicates that the summer monsoon is intensified during the onset. This is consistent with what were found in Qian et al (2011), which proves that our experiments are not contradictory to previous studies on black carbon.

However, the same model with the SDE of dust (not specifically focused) gives a quite different response of monsoon, which indicates different mechanisms behind

SDE of dust and black carbon. As we proposed, the spatial distributions of dust and black carbon are not similar. The main difference is that black carbon from the industrial countries is generally transport eastwards and scarcely into upwind Central Asia. Central Asia is also covered by snow although far less than TP. As a result, the forcing of black carbon is restricted to western TP but the forcing of dust is over Central Asia and western TP. These differences in surface warming by dust and black carbon are also simulated in the experiments by NASA Goddard Earth Observing System Model (Yasunari et al., 2014).

The SDE of dust induces significant warming over western TP and Caspian Sea in Central Asia (Figure 5c), which leads to two cyclonic anomalies over these two regions (Figure 6a). These two cyclonic anomalies intensify the northern branch of Indian monsoon westerly, allowing more dry air from Central Asia penetrating into the monsoon region. But the southern branch of the monsoon westerly is decreased with the associated anticyclonic anomaly over Arabian Sea and India, which weakens the moisture transport from oceans in the south.

There is no doubt that the important role of TP temperature change in Indian monsoon development (Li and Yanai, 1996; Wu et al., 2005; Liu et al., 2001), as the reviewer said. We did not argue against it, however, the role of dust source temperature (not mentioned before) is highlighted from our results. We put Figure R2 in the supplement and largely revised the manuscript (Page 6 Lines 14-17, 21-26; Page 7 Lines 3-8, 13-16, 24-28; Page 8 Lines 4-6, 9-11, 29-35; Page 9 Lines 6-9, 18-29; Page 10 Lines 23-28). Some paragraphs in the original text are deleted. We do not show these intensive revisions here and please see the text.

The dust DRE impacts on Indian summer monsoon is also inconsistent with previous studies, and the results are difficult to understand. The authors found dust DRE could induce a significant cooling over TP, and the cooling is attributed to snow-albedo feedback. The dust AOD is very small over TP (less than 0.05), which implies very weak DRE. How such small DRE produce strong snow-albedo feedback over TP? The feedback processes should be detailed explained. More confusing thing is that the Indian summer monsoon (ISM) is intensified associated with the TP cooling, which is similar to the response induced by TP warming (Lau et al., 2006). The authors simply explained it as a response to downward motion right over TP, which is not convincing. Please provide detailed explanations and supportive reference.

RE: Our black carbon experiments show that the DRE of black carbon can strengthen the monsoon, consistent with Lau et al. (2006). A surface warming over western TP is simulated (Figure R3a). The warming is also over northern India although it is not significant. This effect strengthens the southwesterly winds over the Arabian Sea and moisture transport from ocean (Figure S2b) and the precipitation is intensified over the Arabian Sea and southern India (Figure S2c). These results support that the warmer TP intensifies the monsoon, agreeing with traditional viewpoints (Li and Yanai, 1996; Wu et al., 2005; Liu et al., 2001).

However, comparing the responses to dust and black carbon under the same

model and experiment design, we found that the summer monsoon during the onset is both intensified no matter whether it is warming or cooling over the TP. Since the TP cooling is unlikely to intensify the monsoon (at least few evidence support it). We agree it and ascribe the monsoon strengthening to the warming over Arabian Peninsula/Middle East, which also gains strong support from previous researches (Vinoj et al., 2014; Jin et al., 2014; Solomon et al., 2015).

The Arabian Peninsular warming (Figure 5d) induces a local cyclonic anomaly (Figure 6b). The northern branch of monsoon westerly is remarkably reduced in its intensity across the southern slope of the TP, the Persian Gulf and northern Arabian Peninsula (Figure 6b). The southern branch of Indian monsoon westerly over Arabian Sea is simulated to be stronger, which intensifies the water vapor transport from oceans. Although the TP cooling tends to weaken the monsoon, the Middle East warming overacts and induces a stronger monsoon instead. From an observation study, the heating and intensified high pressure cell over Arabian Peninsula is proved to be an important factor affecting the onset of Indian monsoon (Zhang et al., 2014). Thus, based on the new results, we do not emphasize the role of TP and propose the role of dust source temperature in the revision (Page 6 Lines 18-19, 29-33; Page 7 Lines 8-11, 17-18, 28-30; Page 8 Lines 19-21, 29-35; Page 9 Lines 10-14; Page 10 Lines 9-16, 23-28). Figure R3 is also put in the supplement. Some paragraphs in the original text are deleted. We do not show these intensive revisions here and please see the text.

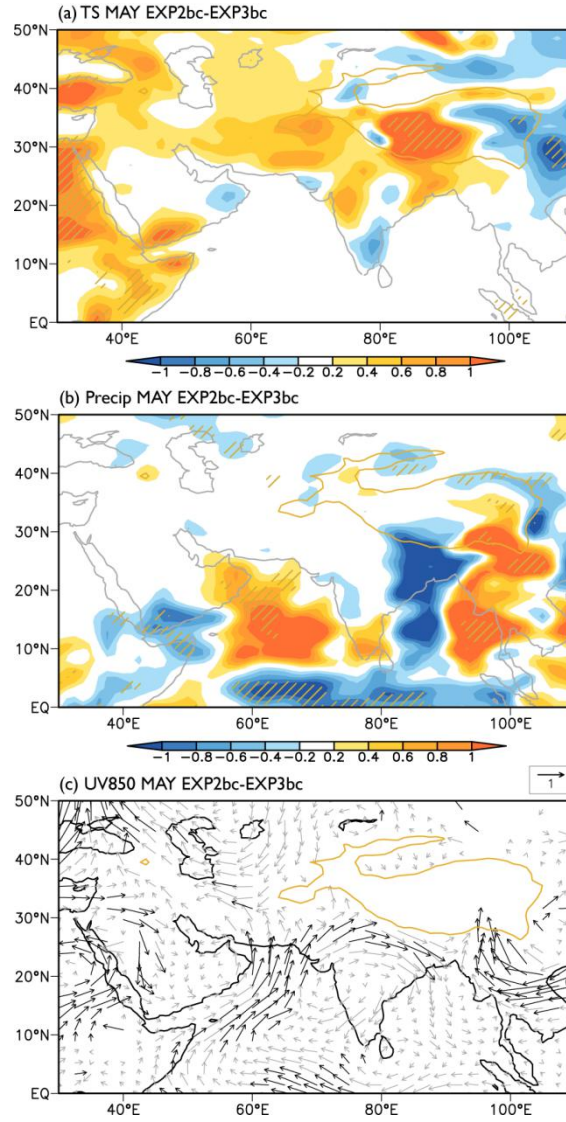


Figure R3: Spatial distribution of changes in precipitation rates (a, mm day<sup>-1</sup>), surface temperature (b) and 850hPa wind vectors (c, m/s) in May induced by direct radiative effect of black carbon.

For the simulated large cooling over TP, we only found in our analysis that the snow albedo feedback amplifies the response of temperature to small dust forcing. Certainly, we can not deny the possible role of other feedbacks. In addition, compare to that of black carbon, the DRE of dust on temperature is largely uncertain (Kok et al., 2017).

In this study, the CAM4 was run with prescribed climatological SST and sea ice. The SST response to aerosol forcing (slow response) is not taken into account. Many previous works showed that the slow response can play a dominant role in the total response of Indian summer monsoon to aerosol forcing (Ganguly et al., 2012). Many previous studies investigated dust impacts on ISM with coupled simulations (e.g. Qian et al, 2011). It could be a possible reason why the monsoon response is opposite to previous studies. Thus, the authors should run coupled simulations and make a

comparison with current results.

RE: Thanks for the comments. We agree with the reviewer that the slow response of ocean may make the response more complicated. However, due to the limited time of final response phase, it is difficult for us to conduct additional coupled model simulations, which are always integrated for hundreds of years for quasi-equilibrium. More importantly, three atmospheric GCM experiments focused on black carbon were conducted to support our arguments. The results strongly support the distinct forcing of black carbon and dust on the Indian monsoon via different mechanisms, as we said in responses above.

In the revision, we cited the references and emphasized the possible role of slow ocean processes (Page 4 Lines 27-29): “Due to the limit of calculation resource, we only conducted atmospheric model experiments in this study and coupled ocean-atmosphere model experiments are not included. Actually, slow ocean response can play a dominant role in the response of Indian summer monsoon to aerosol forcing (Ganguly et al., 2012).”

This study investigates the dust impacts on Indian summer monsoon. However, only the dust effect during the monsoon onset periods (May and June) is investigated. The Indian summer monsoon is from June to August (or September). Please show the monsoon response in July and August, for the dust concentration is still high in Indian at that time (Gu et al., 2016). The response of ISM could be quite different in July and August, for dust DRE impacts could be more important at that time. Only with an examination of the response in entire monsoon period, the title of this manuscript could be appropriate.

RE: Thanks. The response of monsoon during its mature period (July-September) to DRE and SDE of dust is also important. However, these changes in the precipitation and low-high level circulation are similar but complicated, and also not as significant as those during the onset, possibly because the monsoon onset is more sensitive to radiative and temperature changes. Lots of previous studies indicated the sensitive responses of monsoon onset to external forcing, e.g., the sensible heat changes over TP or to its southwest (e.g., Li and Yanai, 1996; Wu and Zhang, 1998; Wu et al., 2012), in agreement with our study. To be accurate, as the reviewer commented, we revised our title to “Indian summer monsoon onset” in the revision.

Other comments:

Page 1, Line 2: “have” should be “has”.

RE: Corrected.

Page 4, Line 6-7: Please give more explanations on “snow-darkening and direct radiative feedbacks”. Does it mean the permit of dust snow-darkening and direct

radiative effects in simulations? What is the meaning of feedback?

RE: Yes. We meant the snow-darkening and direct radiative effects are considered in the experiments. We revised the sentence (Page 4 Lines 10-11).

Page 5, Line 13: Please provide references for the two branches of Indian summer monsoon.

RE: A reference is added here (Wu et al., 2012).

Page 5, Line 25: If EXP1-EXP2 equals to the impacts of dust SDE, please use the dust SDE in the rest of manuscript for consistency. So do the cases for EXP2-EXP3.

RE: We used SDE and DRE instead.

Page 5, Line 26: Please clarify the definition of “Indian monsoon area”.

RE: We specified the region (10-25°N, 65-100°E) in the text (Page 6 Line 1).

Page 5, Line 28: Indian summer monsoon lasts from June to August. Please show the precipitation change in July and August, as well.

RE: As we responded previously, we changed the title to “... Indian summer monsoon onset”. Thus, in the revision, we still showed the monsoon response in May and June.

Page 6, Line 9: Why dust SDE induces significant cooling over Tibetan Plateau? The dust AOD is very small over Tibetan Plateau.

RE: We think here the reviewer means DRE (not SDE). In this paper, the DRE-induced cooling over Tibetan Plateau is explained by the snow-albedo feedback. We do not find the important contributions from other processes in our analyses. More importantly, based on the new results, we do not emphasize the role of TP in the revision. Thus, we turn our eyes on the new-proposed role of dust source temperature and give detailed explanations on this point.

Page 6, Line 15: Why is the southern branch of the monsoon westerly significantly decreased?

RE: We found that surface temperature becomes warmer over most Asia, which responds to the SDE. The most obvious warming is found over western TP where the surface snow cover is larger. Another significant warming center is around Caspian Sea in Central Asia also with certain snow covers at this time. Following the temperature changes, a significant cyclonic anomaly is simulated over western TP and there is also a cyclonic anomaly around the Caspian Sea. These two cyclonic



anomalies tends to intensify the northern branch of Indian monsoon westerly, allowing more dry air from Central Asia penetrating into the monsoon region. However, the southern branch of the monsoon westerly is significantly decreased due to the associated anticyclonic anomaly over Arabian Sea and India.

We emphasized it in the revision (Page 6 Lines 21-26): “In the SDE-induced difference, a significant cyclonic anomaly is simulated over western TP and to its west there is also a cyclonic anomaly around the Caspian Sea (Figure 6a), following the surface temperature changes (Figure 5c). These two cyclonic anomalies tends to intensify the northern branch of Indian monsoon westerly, allowing more dry air from Central Asia penetrating into the monsoon region. However, the southern branch of the monsoon westerly is significantly decreased with the associated anticyclonic anomaly over Arabian Sea and India, which weakens the moisture transport from oceans in the south.”

Page 8, Line 10: Please show the dust snow forcing (outputted by SNICAR), dust deposition (dry and wet), and dust concentration in snow over TP and their seasonal variation. A comparison with previous studies (e.g., Qian et al., 2011) is also needed.

RE: We analysed the suggested variables and the results are shown in Figure R4. It is clearly seen that the dust deposition flux and concentration in top snow layer reach its peak in boreal spring. The forcing (as shown by changes in surface radiation, snow cover fraction and albedo) due to SDE of dust is maximal during April-June, the vital period for snow melting. The peak of dust forcing lags the deposition by about one month, which indicates the memory effect of snow processes. The seasonality of dust deposition and dust snow forcing is similar with previous studies, supporting that our experiments are reasonable. Due to the change of our emphasis, these changes over TP are not added to the revision.

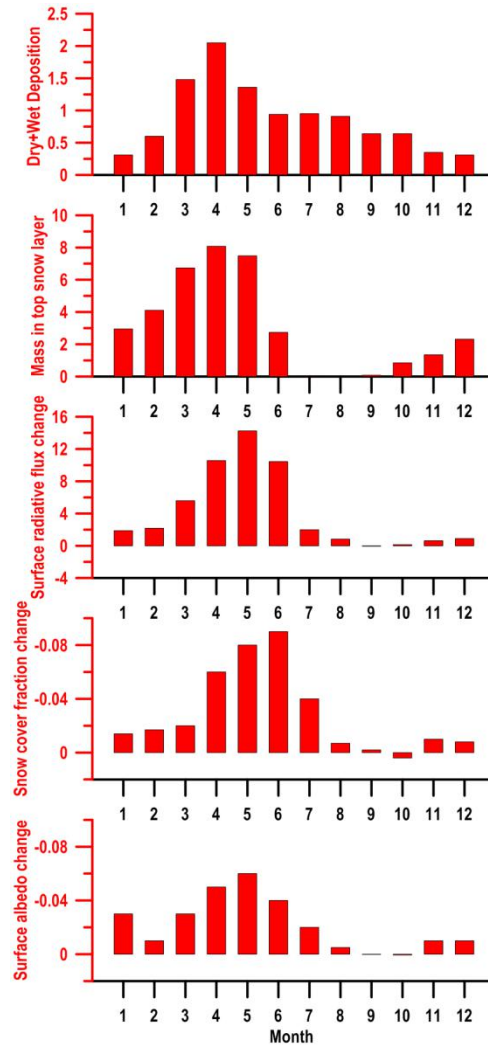


Figure R4: Total dust deposition fluxes and mass of dust in top snow layer in EXP1 experiments and the changes in surface radiative fluxes, snow cover fractions and surface albedo due to SDE of dust averaged for the TP region (70-90°E, 30-45°N)

Page 8, Line 23: Please explain the feedback.

RE: We meant the snow-albedo feedback here based on our analyses.

Page 8, Line 23: Dust aerosols could absorb both shortwave and longwave radiative fluxes. Why the longwave radiative flux change is negative?

RE: From our results, only the net longwave forcing for column atmosphere is negative. The reason is that the warmer atmosphere as a black body emits more longwave radiation, which exceeds over the absorbed amount. Previous studies also showed similar features for net longwave radiation change (Albani et al., 2014; Xie et al., 2018).

Page 8, Line 34: The dust AOD is very small over TP (less than 0.05), which implies

very weak DRE. How such weak DRE produce significant snow cover increase and surface cooling over TP? It could not be simply attributed to snow-albedo feedback.

RE: In this paper, the DRE-induced cooling over Tibetan Plateau is explained by the snow-albedo feedback. We do not find important contributions from other processes in our analyses. More importantly, based on the new results, we do not emphasize the role of TP in the revision. Thus, we turn our eyes on the new-proposed role of dust source temperature and give detailed explanations on this point.

Page 9, Line 19-30: In Lau et al. 2010, they found that TP warming tends to increase the Indian precipitation in May, and the warming center is located at western TP. Their result is consistent with Qian et al. 2010, but different with the results of this manuscript. Explanations are needed here.

RE: Please see Figure R2 and the associated response.

Page 10, Line 12: How could downward motion right over TP induce an upward motion over Indian ? Is it noticed any previous studies? Please provide more explanations as well as the references.

RE: Thanks. We revised this assertion because we do not have enough evidence. As we discussed in other responses, the TP cooling and downward motion may be not closely associated with the intensified monsoon and upward motion over India. In the revision, we ascribed the intensified monsoon to the warming over Arabian Peninsula, which gains strong support from previous studies (Vinoj et al., 2014; Jin et al., 2014; Jin et al., 2015; Solomon et al., 2015).

Figures:

Figure 2 and Figure 3 could be put in the supplement, for they are too many figures for this manuscript.

RE: We kept Figure 2 and Figure 3 in the manuscript because we removed several figures in the revision. We will put them in the supplement if the reviewer still feels there are too many figures.

Figure 4: Please use the specific date in figure 4 (e.g. May 1st).

Figure 4: Please specify the regions of precipitation change.

RE: We changed the date and also specified the region (10-25°N, 65-100°E) in the caption.

Figure 5: Please display the precipitation and surface temperature with different color tables.

RE: We used different color bars.

Figure 5 and so on: Please show “SDE” and “DRE” in figure title.

RE: We used “SDE” and “DRE” instead of “EXP1-EXP2” and “EXP2-EXP3”, respectively.

Figure 5 to Figure 10: There are too many figures for this part. Decide what is important and put the rest in supplement.

RE: We removed original Figure 7, 10, 13 because we wish to avoid the repeating and also do not emphasize the TP temperature any more. Two new figures, showing black carbon’s related results, are added in the supplement.

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# Snow-darkening versus direct radiative effects of mineral dust aerosol on the Indian summer monsoon onset: role of dust source temperature changes

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**Abstract.** Atmospheric absorptive aerosols exert complicated effects on the climate system and two of which are through their direct radiative forcing and snow-darkening forcing. Compared to black carbon, the snow-darkening effect of dust on climate has been scarcely explored till now. When depositing in snow, dust can reduce the albedo of snow by darkening it and increase the snow melt. In this study, the snow-darkening effect of dust, as well as the direct radiative effect, on the Indian summer monsoon are evaluated by atmospheric general circulation model experiments. The results show that, the snow-darkening and direct radiative forcing of dust have both significant impacts on the onset of Indian monsoon but they are distinctly opposite. The snow-darkening effect of dust weakens the Indian monsoon precipitation during May and June, opposite to black carbon. The surface temperature over Central Asia and western Tibetan Plateau becomes warmer due to the dust-induced decrease in snow cover, which leads to a local low-level cyclonic anomaly as well as an anticyclonic anomaly over Indian subcontinent and Arabian Sea. This circulation pattern allows air current penetrating into Indian subcontinent more from Central Asia but less from Indian Ocean. In contrast, the direct radiative forcing of dust warms Arabian Peninsular, which intensifies moisture convergence and precipitation over Indian monsoon region. The upper tropospheric atmospheric circulation over Asia is also sensitive to both effects. Our results highlight an important role of dust source temperature changes in the effects of dust, which significantly modifies the response of summer monsoon. Thus, links between the climatic impact of dust and thermal condition over dust source are of importance and require to be clarified accurately.

## 1 Introduction

Mineral dust, a kind of natural aerosols in the atmosphere, mainly originates from the global deserts including Sahara, Arabian peninsula, Central Asia and East Asia. Dust emitting into the atmosphere is carried by atmospheric circulation and can be transported downwind for a long distance (Zhang et al., 1997; Zhao et al., 2006; Kallos et al., 2006; Schepanski et al., 2009; Shi and Liu, 2011). Mineral dust aerosol affects global and regional energy budget, formation of clouds and precipitation as well as various climate systems through their direct, semi-direct and indirect effects (e.g., Tegen and Lacis, 1996; Ramanathan et al., 2001; Miller et al., 2004; Shao et al., 2011; Huang et al., 2014; Mahowald et al., 2014). Among the climatic effects of dust, the direct radiative effect (DRE) and snow-darkening effect (SDE) are two important components, which exert great impacts on the radiative balance (Haywood et al., 2001; Flanner et al., 2009; Huang et al., 2014; Qian et al., 2015).

The DRE of dust is that the particles can directly scatter and absorb the solar shortwave and black-body longwave radiation. In the fifth Assessment Report (IPCC, 2013), the annual mean DRE of dust is approximately  $-0.10 \text{ W m}^{-2}$  on the global scale, which varies from  $-0.30$  to  $+0.10 \text{ W m}^{-2}$  among different global climate models. However, it is still unclear whether dust aerosol has a net warming or cooling effect on global climate (e.g., Tegen and Lacis, 1996; Miller and Tegen, 1998; Mahowald et al., 2014; Kok et al., 2017; Xie et al., 2018a). Due to the underestimation of coarser dust in climate models than in the atmosphere, the considered DRE may be more cooling in current model ensemble and the possibility that dust causes a net warming is highlighted (Kok et al., 2017).

Following the changes in radiative balance, specific climate systems or atmospheric circulations also respond significantly to the DRE of aerosols. During the emission seasons, dust from inland Asian and Arabian deserts is delivered downwind by the westerlies and Asian monsoon (Uno et al., 2009; Shi and Liu, 2011; Vinoj et al., 2014) to eastern China, India and even deposits in the Tibetan Plateau (Huang et al., 2007; Xu et al., 2009; Zhang et al., 2018). Such distributions of atmospheric dust largely affect the Asian climate, including both Indian and East Asian monsoon (Lau et al., 2006; Zhang et al., 2009; Sun et al., 2012; Vinoj et al., 2014; Jin et al., 2014; Gu et al., 2016; Lau et al., 2017; Lou et al., 2017). Via a strong effect of elevated heat pump, the DRE of absorbing aerosols including dust enhances the heat source over the TP and results in a northward shift of Indian summer monsoon during the late spring and early summer (Lau et al., 2006; Lau et al., 2017).

The aerosol-induced upper tropospheric warming intensifies the updraft air motion, which pumps more moist air from south oceans to north India. However, this hypothesis is still in debate that lacks of observational support (Nigam and Bollasina, 2010). Beside the TP warming, the tropospheric warming over Arabian Sea and surrounding regions due to mineral dust from Middle East can intensify the Indian summer monsoon and precipitation (Vinoj et al., 2014; Jin et al., 2014). In addition, the East Asian monsoon and the eastern precipitation are also significantly affected by dust that northeasterly wind anomaly over eastern China seems to weaken the monsoon circulation (Sun et al., 2012; Tang et al., 2018).

SDE is another important effect of dust on climate, which is not mentioned as intensively as the DRE. Light absorbing aerosols can darken the snow and reduce the surface albedo when deposited in snow, and it can also absorb the radiation and warm the snow surface, which both accelerates the melt process of snowpack (Hansen and Nazarenko, 2004; Xu et al., 2009; Lee et al., 2013; Qian et al., 2015). Due to the reduction of snow, the SDE of absorbing aerosols generally induces a net regional



warming over the snow cover areas. Black carbon, as the most important anthropogenic absorbing aerosols, has a global-mean radiative forcing of +0.04 (+0.02 to +0.09) W m<sup>-2</sup> for SDE (Bond et al., 2013). Over the Tibetan Plateau (TP) where most areas are covered by snow, in particular, the absorbing aerosols in snow remarkably influence the snow albedo and promote the snowmelt (Lau et al., 2010; Yasunari et al., 2011; He et al., 2014; Zhao et al., 2014; Lee et al., 2017; Niu et al., 2017). The SDE of black carbon generates positive changes in surface radiative flux of about 5–25 W m<sup>-2</sup> over the TP during springtime, warms the surface TP obviously and intensifies both the Indian and East Asian summer monsoon by enhancing the elevated heat source (Qian et al., 2011; Qian et al., 2015).

Compared to that of black carbon, the SDE of mineral dust over TP and Asia, especially its influence on the Asian monsoon, is still not clear. Theoretically, the SDE of dust is considered to be larger than that of black carbon over the TP (Flanner et al., 2009; Ming et al. 2013) primarily because the concentration of dust is much higher. The spatial distribution and deposition of dust is also different from black carbon that the dust can be deposited over both central Asia and TP where exists a fraction of snow cover while black carbon is primarily restricted to South and East Asia and downwind areas. In actual, the dust is a kind of natural aerosols, differing from black carbon which is mainly anthropogenic produced. Beside the modern period, the climatic effect of Asian dust are also of great importance in the geological stages, such as the last glacial maximum (Harrison et al., 2001; Claquin et al., 2003; Takemura et al., 2009). During the late Cenozoic, the dust effect ought to become gradually larger as deserts expand and atmospheric dust increases with plateau uplift and climatic cooling (Shi et al., 2011). Thus, it is necessary to explore in detail the effect of dust during present day and geological periods.

In this paper, as a first step, we employed a set of numerical experiments by a general circulation model to evaluate the SDE and DRE of dust on Indian summer monsoon during the onset under present-day conditions. In Section 2, the model and experiments are described. The model performance, response of Indian monsoon and role of temperature changes over dust sources are presented in Section 3. The discussion and conclusions are summarized in Section 4 and 5, respectively.

## 2 Model and Experiments

An atmospheric general circulation model namely Community Atmosphere Model 4 (CAM4), which is improved with a new bulk aerosol model (BAM) parameterization, is employed to evaluate the response of Indian summer monsoon to the forcing of mineral dust. CAM4 is the atmospheric component of the Community Climate System Model 4 (CCSM4), which is coupled with the Community Land Model 4 (CLM4) for land surface processes. The vertically Lagrangian and horizontally Eulerian coordinates are used in the finite-volume discretization of this model. The dust cycle including the emission, transport and deposition, is parameterized in CAM4 and its radiative feedbacks are also calculated on line. The dust sizes in CAM4 contain four bins of 0.1–1.0 μm, 1.0–2.5 μm, 2.5–5.0 μm and 5.0–10.0 μm in diameters, respectively (Mahowald et al., 2006). The CAM4-BAM has been improved by an optimized soil erodibility map and a new size distribution for dust emission (the percentages for four bins are 0.02, 0.09, 0.27, 0.62, respectively), as well as updated optical properties for radiation budget, to present a better performance on simulating the global dust cycle (Albani et al., 2014). In CAM4-BAM, the SDE of all aerosols are enabled but the indirect effect is not considered, which means that the aerosol changes in cloud process as condensation

nuclei are prescribed. Wet removal through in-cloud process is not considered, which may induce bias of dust deposition on snow over Asia. The snow darkening processes are considered based on the Snow, Ice and Aerosol Radiative (SNICAR) module (Flanner et al., 2007; 2009) in which the dust and black carbon aerosols are included. The SNICAR applies Mie scattering to particle mixture and a multi-layer radiative transfer approximation (Toon et al., 1989) to represent vertical inhomogeneity in the snow. The radiative transfer in the snow is affected by the vertical particle profile controlling by fresh snow and flushing with melt water when dust deposits on the surface. Dust optical properties in snow were ranging from 0.88 to 0.99 with decreasing particle size (Flanner et al., 2009). Of note is that SCINAR assumes external mixing between aerosols and spherical snow grains, however, aerosol-snow internal mixing and nonspherical snow shape could significantly affect aerosol-induced snow albedo effects, based on recent studies (Flanner et al., 2012; Liou et al., 2014; Räisänen et al., 2017; He et al., 2018).

Three sensitivity experiments are conducted in this study to evaluate the SDE and DRE of mineral dust. Both the snow-darkening and direct radiative effects of dust are turned on in the experiment namely EXP1d while only the direct radiative feedback is enabled in the experiment of EXP2d. Neither effects are taken into consideration in the third experiment (EXP3d). Thus, the differences in climate responses between EXP1d and EXP2d, and between EXP2d and EXP3d, are denoted as the SDE and DRE of dust, respectively. Of note is that the dust column loading over Asia is slightly larger by the on-line feedbacks when both two effects are enabled, compared to that when DRE is only enabled. However, the bias does not affect our discussion, which will be mentioned later in this work. The reason for the intensified dust cycle over Asia by SDE is analyzed in detail in a parallel study (Xie et al., 2018b). Other species of aerosols except mineral dust are neglected in these experiments to avoid the biases induced by their different spatial distributions in different experiments. In order to compare with previous studies with a main focus on black carbon, three experiments on SDE and DRE of black carbon are also conducted

(EXP1bc, EXP2bc and EXP3bc, respectively) and the design is similar. For these six experiments, the boundary conditions, including the sea surface temperature and greenhouse gas concentrations, are kept as their modern values (The year 2000 AD). The sea surface temperature and sea ice is given from HadOIBI data and the atmospheric CO<sub>2</sub> concentration is set to 367 ppmv.

In these experiments, the horizontal resolution of CAM4-BAM is set to approximately  $0.9^{\circ} \times 1.25^{\circ}$  in latitude and longitude. All the experiments are integrated for a total period of 21 years and the results of the last 15 years are analyzed. Both monthly and daily mean values of variables are outputted to examine the sensitivity of monsoon. The response of Indian monsoon circulation and precipitation during May and June (i.e., the onset) is focused in this study since the monsoon onset is sensitive to external thermal forcing. Due to the limit of calculation resource, we only conducted atmospheric model experiments in this study and coupled ocean-atmosphere model experiments are not included. Actually, slow ocean response can play a dominant role in the response of Indian summer monsoon to aerosol forcing (Ganguly et al., 2012).

## 3 Results

### 3.1 Model validation

Before the examination of monsoon response, the model's ability on simulating the climatology of dust aerosol optical depth (AOD), snow cover and Indian monsoon during May and June in the experiment EXP1d is first evaluated using modern

observation and reanalysis data. The distributions of the AOD and deposition flux of mineral dust in the model over Asia are shown (Figure 1). The maximal values of May-June mean dust AOD are found over the arid and semi-arid regions including the Sahara, Arabian Peninsula, Central Asian and East Asian deserts (Figure 1a). The AOD reaches above 0.2 over major source areas. This simulated pattern is similar with the Cloud-Aerosol Lidar and Infrared Pathfinder Satellite Observation (CALIPSO)-retrieved AOD over the deserts (Figure 1b), which indicates that CAM4-BAM has a good performance on the dust cycle. The simulated absolute values of dust AOD over Arabian Peninsular, southwestern slope of the TP and Taklimakan desert are biased low because the considered dust particles are restricted to less than  $10.0\ \mu\text{m}$  and the dust forcing is underestimated due to less coarser dusts in the current global climate models (Kok et al., 2017). The total deposition fluxes during March-April and May-June (Figures 1c, 1d) show that there are remarkable dust depositions over Asia and adjacent oceans in both periods. In March and April, the dust deposition over East and Central Asian deserts and downwind regions is larger than that in May and June. In contrast, the deposition over Arabian Peninsula is more obvious in May and June, which is also detected over Arabian Sea and western Indian continent. Over the western and northeastern Tibetan Plateau (TP), the deposition flux is simulated with a range of about  $0.02\text{-}0.16\ \text{kg/m}^2/\text{yr}$ .

The simulated snow cover fractions over Asia during May and June show that surface snow exists over Central Asia, East Asia and the whole TP, with largest fractions over western TP (Figure 2a). In Moderate Resolution Imaging Spectroradiometer (MODIS) data, the observed snow cover is found over the same regions that maximal values are located around Caspian Sea, Mongolia, western and southeastern TP (Figure 2b), which is qualitatively consistent with that in the EXP1d simulation. Over the western TP, the MODIS observation presents a fraction larger than 80% but the simulated fraction is smaller. In particular, the model underestimates the elevations of finer-scale mountains and corresponding snow cover fractions due to the coarser resolution, e.g., over the Tianshan mountains. The dust deposition in the surface snow over Asia implies a potential influence on surface snow.

For the Indian monsoon climatology, a feature that the monsoon westerly winds are divided into two branches (e.g., Wu et al., 2012), with the northern one from Central Asian dry regions and southern one from moist Indian Ocean, is simulated in the 850hPa winds during May and June (Figure 3a). During the monsoon onset, the southerly winds over this region gradually develop from the south to the north. During the same period, the Indian monsoon precipitation is mainly produced over the western sides of the Indian and Indo-China peninsulas as well as the southern slope of the TP (Figure 3c). These features of Indian monsoon circulation and precipitation are generally in agreement with the National Centers for Environmental Prediction/National Center for Atmospheric Research (NCEP/NCAR) reanalysis and Tropical Rainfall Measuring Mission (TRMM) satellite-retrieved data (Figures 3b, 3d). Compared to the observations, the simulated precipitation is lighter over the western sides of two peninsulas but heavier over the southern slope of the TP. In brief, CAM4-BAM performs well in both the monsoon climatology and dust cycle over Asia, which builds confidence for assessing the climate sensitivity to dust forcing.

### 3.2 Response of Indian monsoon

The daily precipitation differences during May and June between EXP1d and EXP2d, as well as between EXP2d and EXP3d, are calculated to examine the responses of monsoon onset to SDE and DRE of dust (Figure 4). It is clearly seen that in all

three experiments the precipitation rates over Indian monsoon area (10-25°N, 65-100°E) increase abruptly by an amount of approximately 10 mm day<sup>-1</sup> during several weeks in the onset (Figure 4a). In this two-month period, the dust SDE-induced difference is mostly negative while the dust DRE-induced difference is positive (Figure 4b), which means that the SDE tends to weaken the Indian summer monsoon but the DRE likes to intensify it. This is also the reason why we choose May and June as the monsoon onset in the following analysis. The SDE-induced precipitation decrease exceeds the DRE-induced increase in June, which results in a net reduction in precipitation; however, these two effects almost counteracts by each other and the total precipitation change in May is not significant.

The spatial distributions of May-June mean precipitation show that the precipitation rate is decreased by the SDE over most Indian monsoon regions and a remarkable difference by 1 mm day<sup>-1</sup> is detected over India (Figure 5a). Other regions with statistically-significant precipitation changes are found over western and southeastern TP, parts of Central Asia and northeastern Africa. For DRE-induced response, the precipitation is promoted over Indian peninsula, Arabian Sea and Central Asia but suppressed over Bay of Bengal and southeastern TP (Figure 5b). Thus, the responses of Indian monsoon precipitation to the SDE and DRE are distinctly different during the onset, which highlights the complicated influence of mineral dust. The surface temperature becomes warmer over most Asia, which responds to the SDE (Figure 5c). The most obvious warming, with an amplitude of larger than 1°C, is found over the whole western TP where the surface snow cover is larger, which indicates that the SDE is significant at these regions. Another significant warming center is around Caspian Sea in Central Asia also with certain snow covers at this time. In contrast, the surface temperature difference induced by the DRE is significantly negative over the whole TP and northeastern India (Figure 5d). However, it is simulated to be warming over Arabian Peninsula/Middle East, which amplifies the zonal thermal gradient over Indian monsoon region.

The responses of Indian monsoon circulation to the SDE and DRE of dust are examined by the differences in 850 hPa wind vectors between experiments (Figures 6a, 6b). In the SDE-induced difference, a significant cyclonic anomaly is simulated over western TP and to its west there is also a cyclonic anomaly around the Caspian Sea (Figure 6a), following the surface temperature changes (Figure 5c). These two cyclonic anomalies tends to intensify the northern branch of Indian monsoon westerly, allowing more dry air from Central Asia penetrating into the monsoon region. However, the southern branch of the monsoon westerly is significantly decreased with the associated anticyclonic anomaly over Arabian Sea and India, which weakens the moisture transport from oceans in the south. This circulation anomaly over monsoon area agrees well with the simulated lighter precipitation, which supports that the Indian summer monsoon is weakened by the SDE during its onset. In addition, the westerly winds become stronger to the north of the TP, which might affect the dust emission further over that region. In the DRE-induced difference, the situation is quite different that an cyclonic anomaly is simulated over the Arabian Peninsula, in response to the warming (Figure 5d). The northern branch of monsoon westerly is remarkably reduced in its intensity across the southern slope of the TP, the Persian Gulf and northern Arabian Peninsula (Figure 6b). The southern branch of Indian monsoon westerly over the Arabian Sea is simulated to be stronger, which intensifies the water vapor transport from oceans. The westerly winds are also decreased over the Bay of Bengal and Indo-China Peninsula, however, it brings water vapor to the Indian Peninsula. . The differences in the moisture convergence induced by the SDE and DRE show that the water

vapors diverge and converge over most Indian monsoon region, respectively (Figures 6c, 6d), consistent with the responses of precipitation (Figures 5a, 5b).

Compared to the dust, the SDE and DRE of black carbon on the Indian summer monsoon onset are also analyzed (Figure S1, S2). The black carbon's SDE-induced surface temperature change is similar with dust but is restricted to western TP with no warming over Central Asia (Figure S1a). The TP warming intensifies the Indian monsoon and leads to strong anomaly in southerly winds over India, which subsequently brings more rainfall over this region (Figure S1b, S1c). The SDE of black carbon is distinctly different with that of dust, which indicates the complicated SDE of absorbing aerosols on Indian monsoon not mentioned before. For the DRE of black carbon, a surface warming over western TP is simulated (Figure S2a). The warming is also over northern India although it is not significant. This effect strengthens the southwesterly winds over the Arabian Sea and moisture transport from ocean (Figure S2b) and the precipitation is intensified over the Arabian Sea and southern India (Figure S2c). Thus, the SDE and DRE of black carbon are consistently to intensify the Indian monsoon during the onset.

The responses of Indian monsoon in high-troposphere is examined (Figures 7) because the anomalous heating center over the TP as well as the high pressure cell are both important for the monsoon development. As seen in the 200 hPa climatology, a SDE-induced dipole pattern of meridional temperature changes over Central Asia and TP (Figure 7a) results in a western weakening and a eastern strengthening of South Asian high pressure cell, i.e., a eastward shift of high pressure cell (Figure 7c). In contrast, the opposite dipole temperature changes caused by DRE make the high pressure cell move westward (Figures 7b, 7d). The strong Arabian Peninsular warming over high troposphere is in agreement with the surface (Figure 5d), which indicates that this warming is significant throughout the column atmosphere. Differences in 200 hPa wind vectors also show a couple of reversed circulation changes of cyclonic/anticyclonic cell in the west and anticyclonic/cyclonic cell in the east responding to the SDE and DRE, respectively (Figures 7e, 7f), consistent with the temperature and pressure changes. To the north of the TP, the westerly winds are weakened by the SDE, which might help the long-distance transport of mineral dust over East Asia.

Changes in vertical motion show that low and middle tropospheric subsidence occurs over most of the monsoon areas with the SDE but the DRE leads to ascending motion over Arabian Sea and western India (Figures 8). Strong ascending motion due to the SDE is found over the TP and Caspian Sea (Figure 8a, 8c), which is closely linked with local surface warming (Figure 5c). In contrast, the subsidence dominates the adjacent areas outside the TP including the Indian and Indo-China peninsulas, as well as regions to the west and north of the TP (Figure 8a, 8c), which is in good agreement with low-level circulation changes (Figure 6a). For the DRE, the ascending motion is presented over northern India although surface cooling produces local subsidence over the TP (Figure 8d). The spatial distributions of anomalous vertical motion over Indian monsoon region are in qualitatively coincidence with the simulated precipitation changes by SDE and DRE, respectively. Such circulation changes is also clearly seen in the cross sections for vertical versus meridional winds (not shown).

The SDE on atmospheric radiation budget is mainly realized by surface albedo change led by darkening the snow and accelerating the snowmelt as a kind of absorbing aerosols. A positive feedback loop of SDE is that the net incoming solar shortwave radiation flux increases at the surface when snow albedo is reduced, and rising temperature further enhances the snowmelt and reduces the surface albedo. The SDE-induced differences in longwave and shortwave radiation fluxes for all-sky

conditions during May and June at the top of atmosphere (TOA), at the surface and in the column atmosphere is shown in Figure 9, respectively.

For both the TOA and the surface, the primary forcing of SDE is via shortwave radiation change since it is albedo-induced. Due to large snow cover, the strongest shortwave radiation change is found positive over western TP, Mongolia and the south of Black Sea (Figures 9b, 9e), which indicates that both the TOA and the surface receive more shortwave radiation while the scattering becomes less. The positive shortwave forcing near Indian Peninsula, not so strong as that over TP, is offset by the negative longwave one (Figures 9a, 9d), in which these changes should be associated with internal adjustment of climate, e.g., the water vapor change. As a result, the SDE totally means a positive net radiative forcing over western TP at the TOA and the surface (Figures 9c, 9f), which is the reason for local surface warming (Figure 5c). Additionally, the net surface and column radiative forcing is also positive and statistically significant to the south of Caspian Sea although its absolute value is not as large as that over western TP (Figure 9f, 9i). For the column atmosphere, the shortwave radiation flux does not vary, supporting that the slight dust loading difference between EXP1d and EXP2d merely presents negligible radiation changes. The negative longwave radiation difference is merely found near Indian Peninsula (Figures 9g, 9h, 9i), indicating that the atmosphere loses energy over this region.

For the DRE, the radiative forcing is characterized by positive longwave and negative shortwave radiation differences at both the TOA and the surface (Figures 10a, 10b, 10d, 10e) owing to the absorbing and scattering of radiation by dust. However, the TOA changes are less evident than the surface changes because the dust aerosol is primarily distributed in the low level. Notably, a significant difference of larger than  $20\text{W/m}^2$  in shortwave radiation is seen over western TP (Figures 10b, 10e), highlighting potential feedback of snow albedo. The net surface forcing is also positive over Arabian Peninsula but not statistically significant. Further, the positive net TOA forcing is obvious over Arabian Peninsula (Figures 10c, 10f), which indicates that the pattern of surface temperature change by the DRE (Figure 5d) is more likely controlled by the TOA radiation change. The net TOA and surface radiative forcing of DRE shares a distinct opposite pattern with that of SDE, which is responsible for different response of atmospheric temperature and Indian monsoon. As absorbing aerosol, the longwave and shortwave forcing for the column atmosphere is negative and positive, respectively (Figures 10g, 10h), with maximal values distributed over the large dust AOD region (Figure 1a). The positive net total forcing of dust is found remarkable over Arabian Peninsula but not so large over East Asia (Figure 10i).

From the analysis above, in brief, the suppressed and increased monsoon precipitation during May and June are fundamentally resulted from the SDE and DRE induced changes in atmospheric temperature structure, respectively, especially over the low-level atmosphere where most mineral dust exists. Compared to black carbon, the SDE effect of dust is opposite because the range for dust-induced temperature increase does not only occupies over western TP but also expands to central Asia, which indicates the role of central Asian temperature changes in modulation of SDE on the monsoon. Although the DRE of dust is similar with that of black carbon, the simulated surface temperature changes over the TP are distinctly different, implying that the TP temperature might be not so important, at least in our sensitivity runs. The intensified monsoon is more likely ascribed to the warming over Arabian Peninsula. Thus, the potential importance of temperature changes over dust source (Central Asia and Arabian Peninsula) is highlighted in both SDE and DRE of dust.



## 4 Discussion

The physical mechanisms for SDE and DRE of mineral dust on the Indian summer monsoon during the onset are summarized by schematic diagrams, respectively (Figure 11). The initial forcing of SDE occurs over western TP and Central Asia, which becomes warmer due to the darkened and decreased snow cover. Subsequently, two anomalous surface low pressure centers are produced and upward air flow dominates over these areas (Figure 11a). To their south, a forced high pressure and anticyclone anomaly is found over Bay of Bengal and India where the subsidence suppresses the formation of monsoon rainfall. For the circulation, the anticyclone strengthens the westerly air flow from the dry Central Asia but limits that from moist Indian Ocean. In contrast, the forcing of DRE induce a warming over Arabian Peninsula, which produce a surface low pressure anomaly (Figure 11b). Such a pattern gives a SDE-opposite impact of circulation, which intensifies the cross-equatorial southerly and weakens the dry air flows from the north. The role of western TP cooling and high pressure anomaly is not certain considering that they are different from black carbon-induced changes.

The radiative forcing and remarkable TP warming at the surface and high troposphere, as a direct response to SDE of dust or other absorbing aerosols (e.g., black carbon), is also found in previous studies (Flanner et al., 2009; Lau et al., 2010; Qian et al., 2011). They proposed that the snowmelt process is rapid and efficient during the late spring and early summer (Lau et al., 2010; Qian et al., 2011; Qian et al., 2015) and this sensitive response of snow cover to SDE in melting season supports its significant role in Indian monsoon development simulated in this study. Change in thermal condition over surface TP, which acts as a heat source and exerts great sensible heat flux to atmosphere, are proved to be essential in the establishment of Indian monsoon (e.g., Yanai et al., 1992; Li and Yanai, 1996; Liu et al., 2001). Furthermore, change in snow cover over TP can also obviously affect the Indian monsoon by modifying the thermal TP forcing (e.g., Vernekar et al., 1994; Senan et al., 2016) and those over different parts of TP may play different roles (Wang et al., 2017). The response of Indian monsoon to SDE of black carbon during the onset (Qian et al., 2011, hereafter Qian2011) is similar from what we found here for black carbon. As a result, it seems that the SDE of dust in this study is reasonable although it is different from black carbon. Qian2011 emphasized that the polluted snowpack by black carbon over the TP warms the local surface and enhances the sensible heat flux, which results in a earlier onset of Indian monsoon and heavier precipitation over northern India. The opposite monsoon response to dust originate from different locations of surface warming that the warming due to black carbon is just over the TP but the warming due to dust extends quite westward to Central Asia. The westward extension of warming forces the southerly winds over India (See Figure S1 in this study and Figure 15c in Qian2011) to Arabian Peninsula (Figure 6a). Although the forcing on radiative budget of absorbing aerosols are similar, their distributions obviously differ. Black carbon, mainly emitted from the Industrial countries, is generally transport eastwards and scarcely into upwind Central Asia. These differences in surface warming by dust and black carbon are also simulated in the experiments by NASA Goddard Earth Observing System Model (Yasunari et al., 2014).

The DRE-strengthened Indian summer monsoon in this paper is in qualitatively agreement with previous studies (Lau et al., 2006, hereafter Lau2006; Gu et al., 2016; Lau et al., 2017), in which either dust or black carbon, or both of them, is included. However, the results may share different mechanisms. For example, the DRE-strengthened Indian summer monsoon

in Lau2006 by both dust and black carbon is ascribed to an elevated heat pump (EHP) mechanism that the aerosols heat the southern slope of TP by absorbing the radiation and the hot air rises, which draws in moisture convergence over India. In our black carbon experiments, consistent response of Indian monsoon is simulated although the 3D distribution of black carbon is different. However, the EHP mechanism fails to be obvious when only mineral dust is considered here, because the DRE of dust only induces a remarkable surface cooling over TP during May and June not a warming as shown in our black carbon experiment and Lau2006. The TP warming in Lau2006 is more likely induced by black carbon and we do not make sure whether it can be produced by dust only. Interestingly, there are consistent intensified summer monsoon and upward air motion over northern India in these studies (Lau et al., 2006; Gu et al, 2016). From this perspective, the TP temperature changes might be not important in the DRE of dust. In this study, the intensified summer monsoon due to dust is from Arabian warming over the surface and upper atmosphere (Figures 5d, 7b), which acts like a feedback on dust-monsoon interaction and drives moisture from southern oceans to Indian monsoon areas. This mechanism gains support from previous researches (Vinoj et al., 2014; Jin et al., 2014; Solmon et al., 2015), which emphasized the modulation of western African dust on Indian monsoon rainfall. From an observation study, the heating and intensified high pressure cell over Arabian Peninsula is proved to be an important factor affecting the onset of Indian monsoon (Zhang et al., 2014). Worthy of being pointed out is that the DRE of dust on surface temperature is largely uncertain and depends closely on the size distributions, optical properties and etc (Kok et al., 2017), which restricts our accurate understanding of dust effect. The different performance of dust-induced radiative forcing and temperature changes over East Asia and northern Africa can be explained by different surface albedo background and particle sizes (Liu et al., 2008; Takemura et al., 2009; Su and Toon, 2011; Xie et al., 2018a).

## 5 Conclusions

In this study, significant responses of Indian summer monsoon, including both circulation and precipitation during the onset, are proposed to the SDE and DRE of mineral dust, which is closely associated with the temperature changes over dust sources. The SDE and DRE of dust are found to exert different impacts on monsoon system due to distinct temperature changes over Asia, highlighting the complexity of climate effect of dust. The forcing mechanisms of dust effect and the responses of Indian monsoon are also different from black carbon. Compared to black carbon-induced warming over only western TP, the SDE of dust warms the surface over both Central Asia and western TP, which weakens the monsoon development and precipitation during May and June. The DRE of dust warms Arabian Peninsula and intensifies the monsoon onset. As net result of SDE and DRE of dust, the precipitation in June is reduced. Beside the Indian monsoon, East Asian monsoon should be also affected by the dust-induced thermal change, which will be examined in future. Compared to black carbon presenting positive TOA forcing, the DRE of dust on atmospheric radiation budget and thermal structure are still uncertain, which adds difficulty to evaluate the sensitivity of specific climate system to dust effect. Nevertheless, the role of dust still requires to be deeply explored due that it is natural and ought to be important during past climate change. In particular, several times larger dust burden and deposition during the Last Glacial Maximum (Mahowald et al., 2006; Maher et al., 2010), as well as higher snow cover fraction due to cold climate, are likely to induce stronger DRE and SDE than present day.

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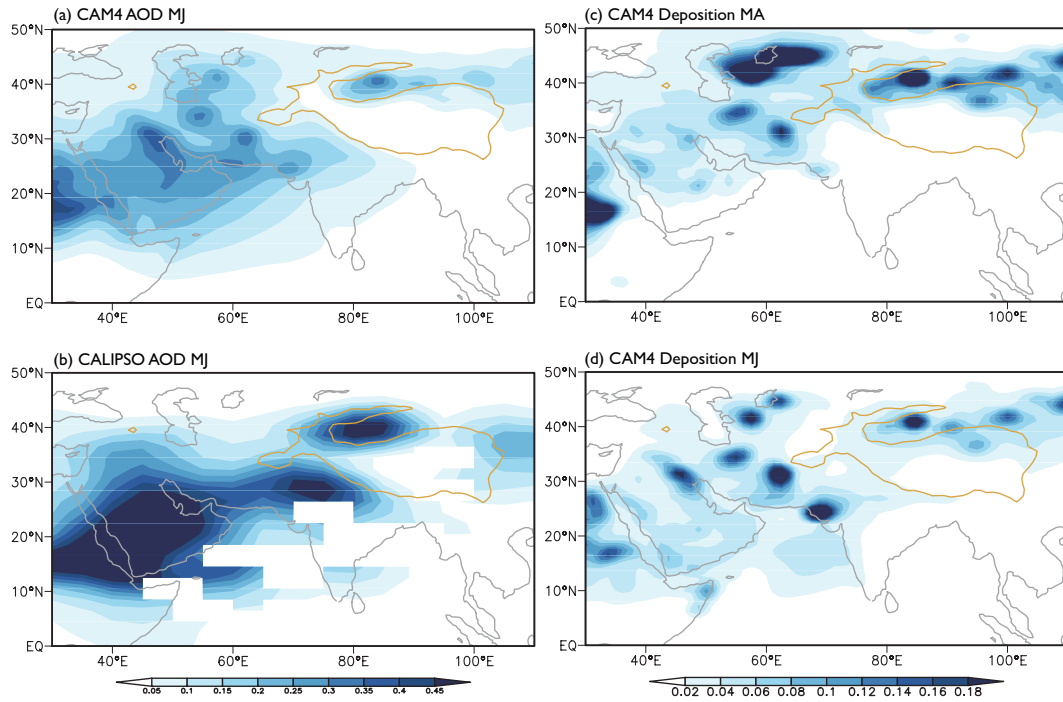
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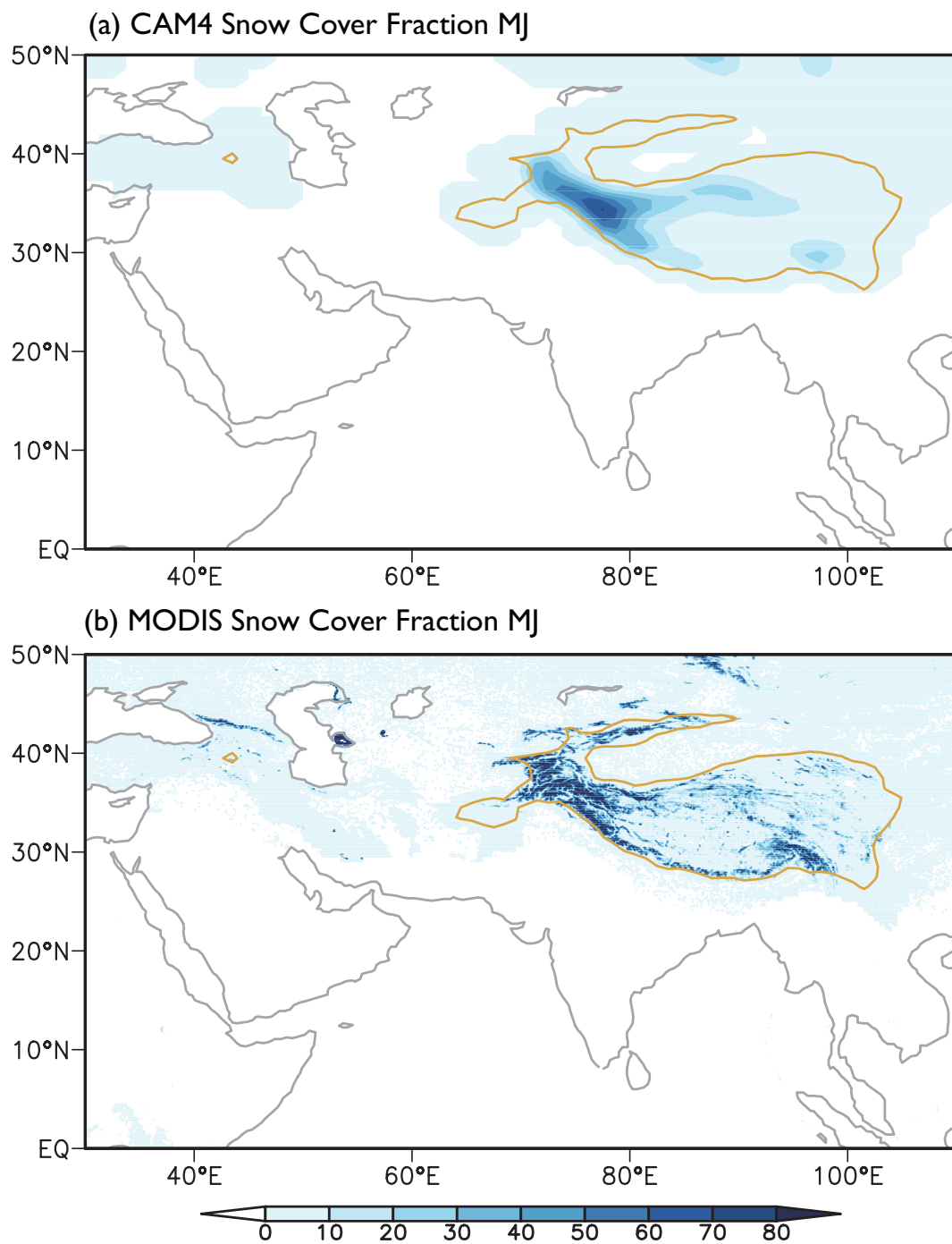
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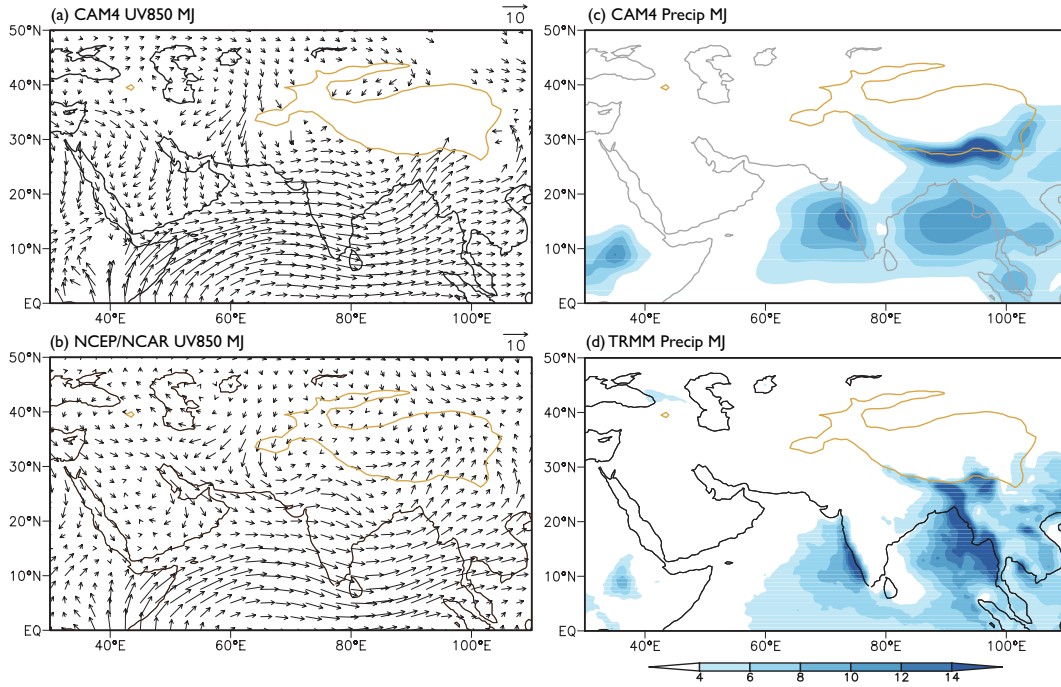
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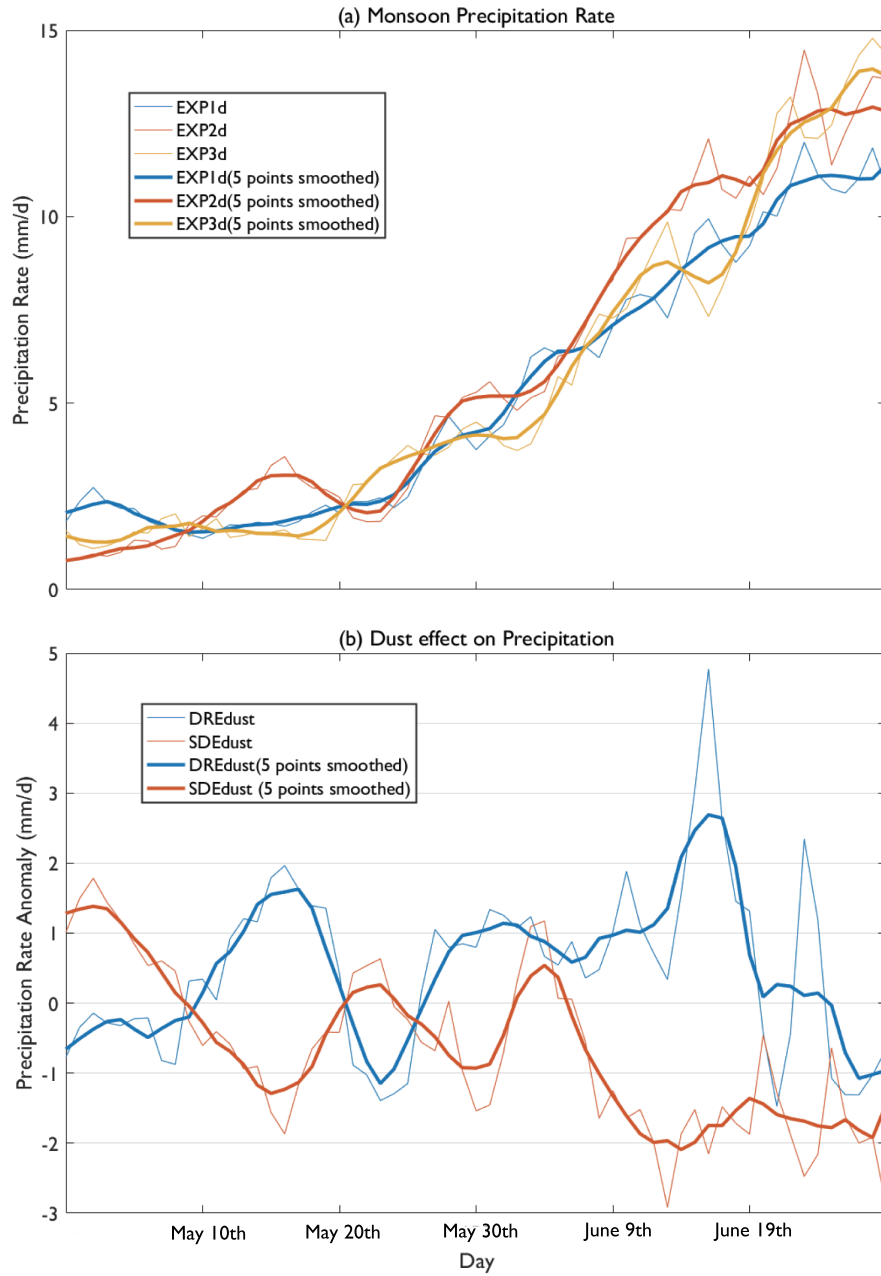
**Figure 1.** Averaged dust aerosol optical depth over Asia for May and June in CAM4 (a) and in Cloud-Aerosol Lidar and Infrared Pathfinder Satellite Observation (CALIPSO)-retrieved data for 2007-2011 (b); and mean dust deposition flux including both dry and wet deposition for March and April (c) (kg/m<sup>2</sup>/yr) and for May and June (d). Yellow line shows the profile of Tibetan Plateau above 2500 m.



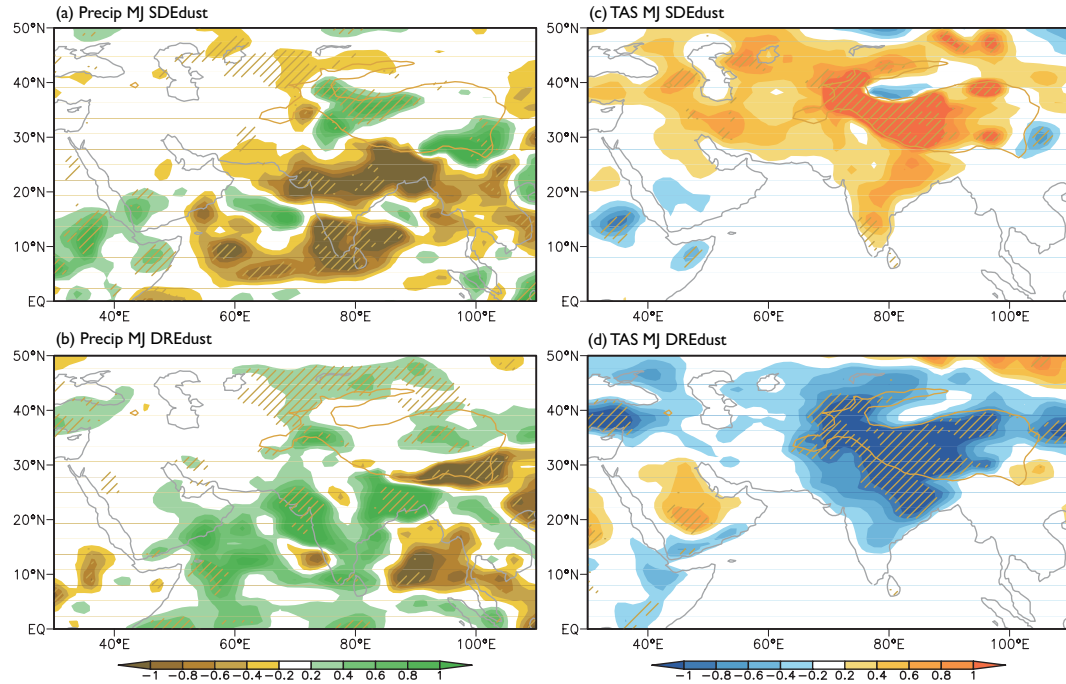
**Figure 2.** Snow cover fraction (%) over Asia for May and June in CAM4 (a) and in Moderate Resolution Imaging Spectroradiometer (MODIS)-retrieved observation. Yellow line shows the profile of Tibetan Plateau above 2500 m.



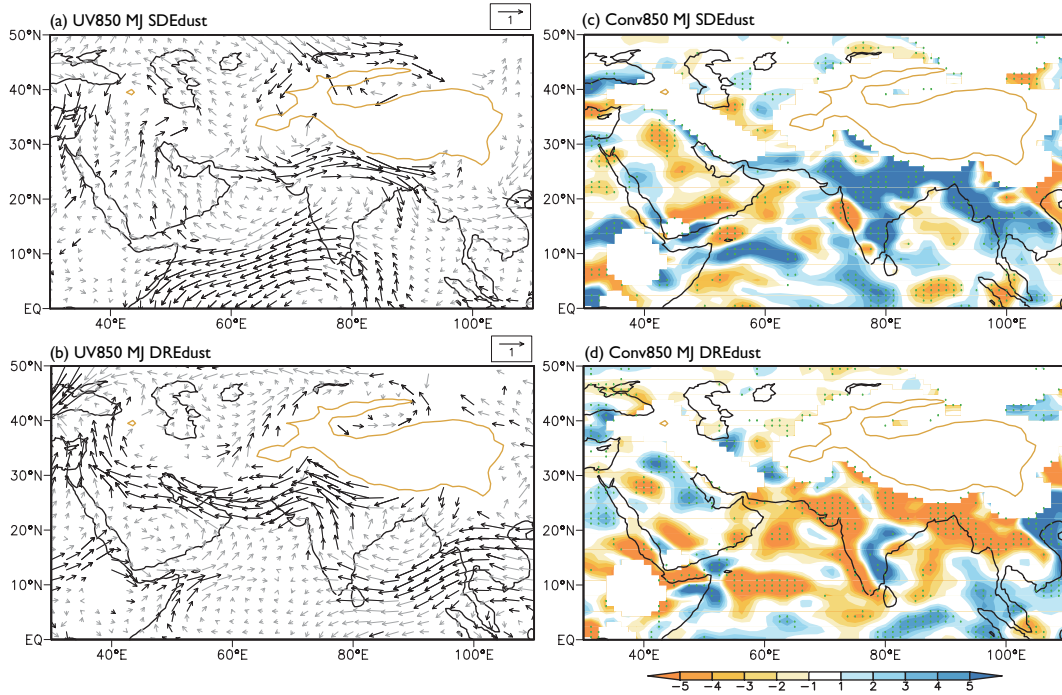
**Figure 3.** Averaged 850 hPa wind vectors ( $\text{m s}^{-1}$ ) over Indian monsoon region for May and June in CAM4 (a) and in National Centers for Environmental Prediction/National Center for Atmospheric Research (NCEP/NCAR) reanalysis data (b); and precipitation rates ( $\text{mm day}^{-1}$ ) for May and June in CAM4 (c) and in Tropical Rainfall Measuring Mission (TRMM)-retrieved data (d). Yellow line shows the profile of Tibetan Plateau above 2500 m



**Figure 4.** Daily precipitation rates ( $\text{mm day}^{-1}$ ), averaged for monsoon area ( $10\text{-}25^{\circ}\text{N}$ ,  $65\text{-}100^{\circ}\text{E}$ ), during May and June in three experiments (a) and the differences ( $\text{mm day}^{-1}$ ) induced by snow-darkening effect and direct radiative effect of dust (b). Thin lines show the daily values and thick ones are 5-day smoothed. In b, red lines donate snow-darkening effect and blue lines donate direct radiative effect.

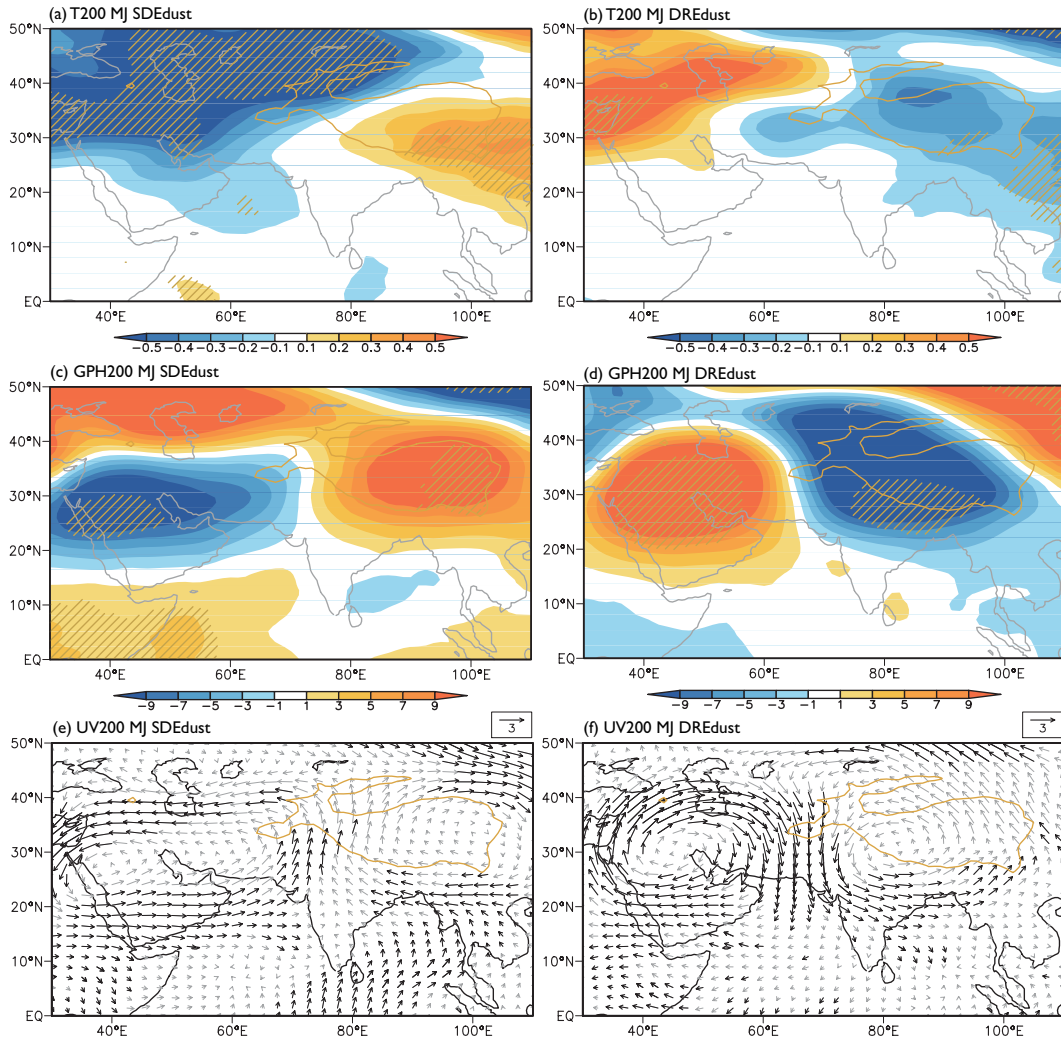


**Figure 5.** Spatial distribution of changes in precipitation rates (a, b,  $\text{mm day}^{-1}$ ) and surface temperature (c, d,  $^{\circ}\text{C}$ ) in May and June induced by snow-darkening effect (top) and direct radiative effect of dust (bottom), respectively. Oblique lines indicate differences significant at 95% confidence level. Yellow line shows the profile of Tibetan Plateau above 2500 m.

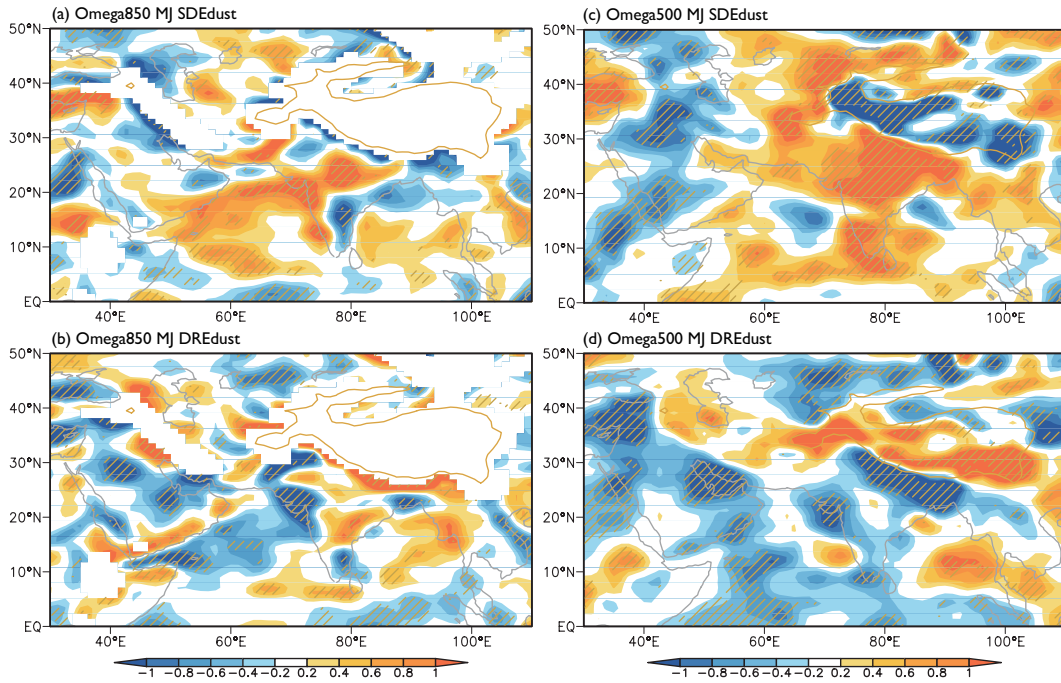


**Figure 6.** Spatial distribution of changes in 850 hPa wind vectors (a, b,  $\text{m s}^{-1}$ ) and moisture convergence (c, d,  $\text{g s kg}^{-1} \text{m}^{-1}$ ) in May and June induced by snow-darkening effect (top) and direct radiative effect of dust (bottom), respectively. Positive values in c and d means divergence anomaly and negative means convergence. Black arrows and green dots indicate differences significant at 90% confidence level. Yellow line shows the profile of Tibetan Plateau above 2500 m.

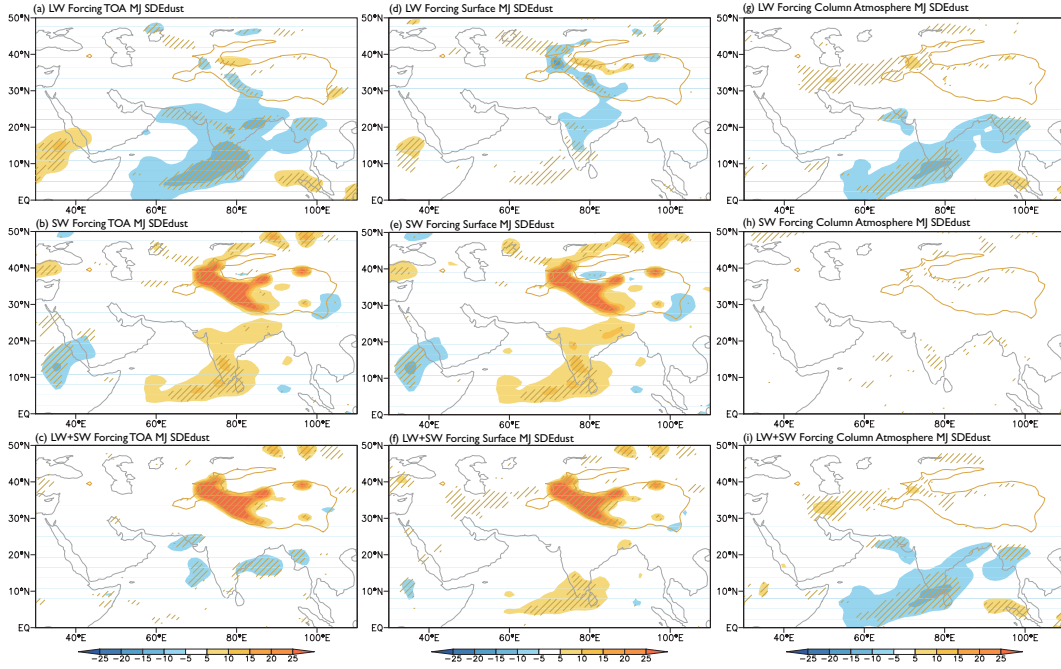




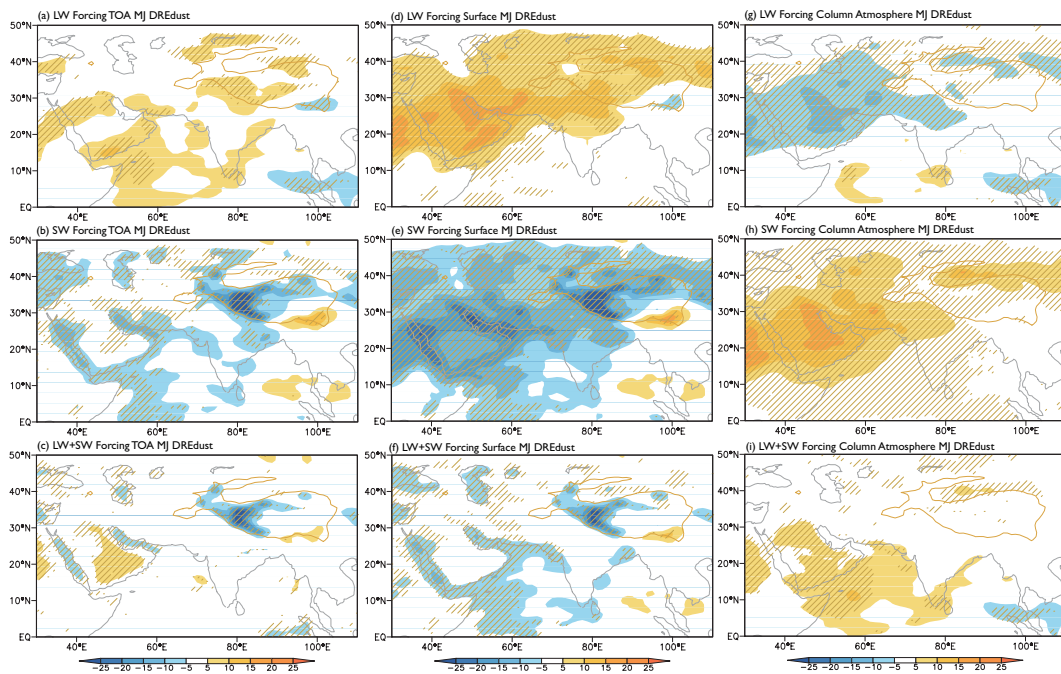
**Figure 7.** Spatial distribution of changes in 200 hPa temperature (a, b,  $^{\circ}\text{C}$ ), geopotential height (c, d, gpm) and wind vectors (e, f,  $\text{m s}^{-1}$ ) in May and June induced by snow-darkening effect (left) and direct radiative effect of dust (right), respectively. Oblique lines and black arrows indicate differences significant at 90% confidence level. Yellow line shows the profile of Tibetan Plateau above 2500 m.



**Figure 8.** Spatial distribution of changes in 850 hPa (a, b) and 500 hPa (c, d) vertical wind speed ( $\times 100, \text{Pa s}^{-1}$ ) in May and June induced by snow-darkening effect (top) and direct radiative effect of dust (bottom), respectively. Negative values indicate upward flow and positive indicate downward flow. Oblique lines indicate differences significant at 95% confidence level. Yellow line shows the profile of Tibetan Plateau above 2500 m.

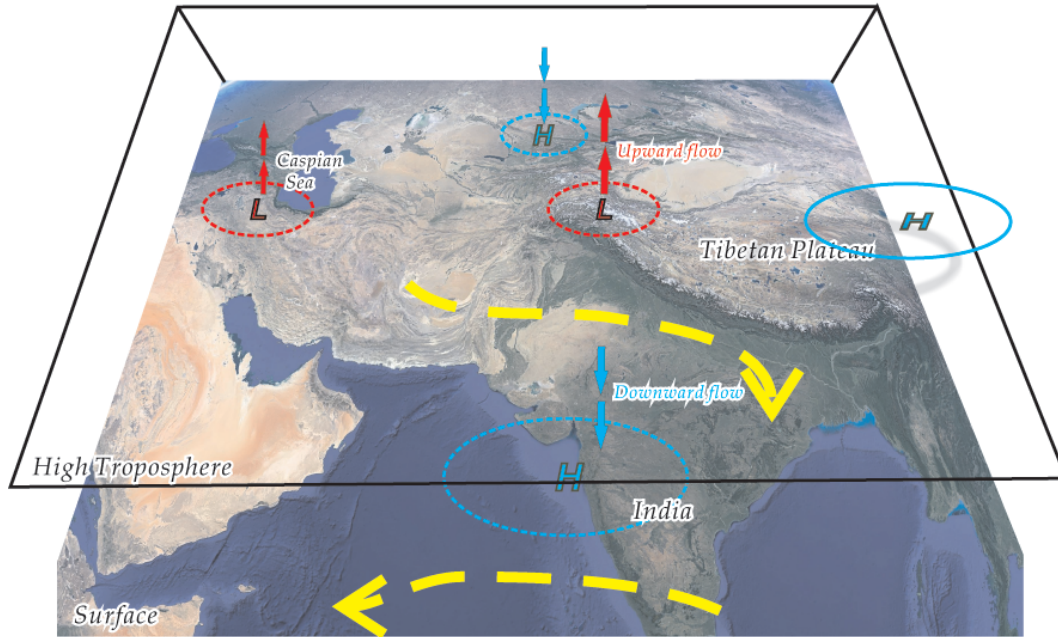


**Figure 9.** Changes in longwave (top,  $\text{W m}^{-2}$ ), shortwave (middle,  $\text{W m}^{-2}$ ) and net (longwave+shortwave, bottom,  $\text{W m}^{-2}$ ) radiative fluxes during May and June by snow-darkening effect of dust for the top of atmosphere (TOA, a-c), the surface (d-f) and the column atmosphere (g-i). Oblique lines indicate differences significant at 95% confidence level. Yellow line shows the profile of Tibetan Plateau above 2500 m.

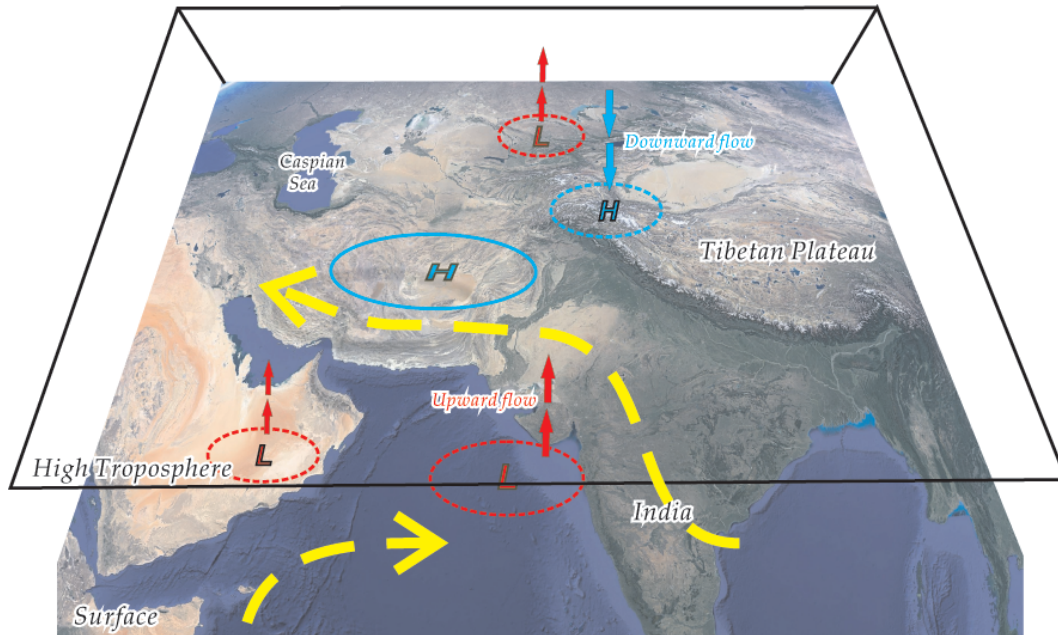


**Figure 10.** Similar with Figure 11, but for direct radiative effect of dust.

(a) Snow-Darkening Effect



(b) Direct Radiative Effect



**Figure 11.** Schematic diagrams showing the forcing mechanisms of snow-darkening effect (a) and direct radiative effect (b) of mineral dust on Indian monsoon during the onset. The circles and abbreviations in them denote the anomalous pressure centers: high pressure (blue), low pressure (red), near surface (dashed) and high troposphere (solid). The red and blue arrows indicate the upward and downward air flows, respectively, and the yellow ones present the differences in horizontal winds.